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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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ORIGIN OF PLAYA STONE TRACKS, RACETRACK PLAYA, INYO COUNTY, CALIFORNIA

BY GEORGE M. STANLEY

ABSTRACT

Curious tracks on playas, obviously made by moving stones, have been interpreted as caused by wind blowing stones over wet level clay. The stones weigh from ounces to several hundred pounds.

Among hundreds of tracks on Racetrack Playa, Inyo County, California, certain ones exhibit near parallelism which implies unit movement; precise measurements and plots confirm this. Certain bends and cusps in these irregular tracks are comparable in all tracks of the same "signature". A transparency plot of analogous points (one for each track in same signature group), when moved along plotted tracks, matches repeatedly at other analogous points. Rotation precluded true parallelism of tracks and identity of lengths and shapes. Distance of unit movement exceeded 300 feet, and maximum spread of stones in unit is 480 feet.

Unit movement over so great a span scarcely allows any reasonable conclusion as to cause other than wind-blown ice floes dragging protruding stones. Ice ramparts and other evidence indicate longshore shearing motion, feasible for ice floes but impossible for ice shove by thermal expansion.

The writer finds no evidence that stones, freely wind blown, have made tracks. Though not disproved, this idea meets serious objections in many other small object tracks than those surveyed and is unfeasible for tracks inscribed by 300 pound stones.

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Introduction

Playa stone tracks have raised the question of whether wind can move innumerable loose rocks ranging from a few ounces to hundreds of pounds over the flat surfaces of playas. Such was originally proposed, and several writers have furthered this supposition.

Many rocks have traveled considerable distances over playa clay. Racetrack Playa (Fig. 1) in the Death Valley region of California is replete with examples; several considerations eliminate the possibility of human agency accomplishing all these, as is agreed by everyone who has studied them. Plate 1 illustrates a typical track 62 feet long, and the 50–60-pound stone that made it. Many such tracks extend several hundreds of feet. Many traveled stones are larger. As stated by Kirk (1952b, p. 177) for the southeast part of Racetrack Playa, almost half a mile due north of its southern extremity, "five large pieces of limestone ranging in size from 200 to 600 pounds, each with a trail, were observed." Thousands of much smaller stones have made tracks some of which manifested their origin and prompted mapping them and writing this paper (Pl. 2).

The writer's first visit to the Racetrack was an incidental, 1-hour inspection of a few stone trails in late December 1951. Skepticism on the wind theory (motivating force applied directly by wind) stimulated another examination, for a day and a half, in April 1952. Two or three groups of tracks then observed displayed a striking intragroup parallelism and favored an alternate interpretation of origin by wind-blown ice floe.

Sketch notes, azimuths, and photos (Pl. 2, figs. 2, 4) were made of the *N* tracks, but were insufficient to test this theory. Detailed mapping (Figs. 2–4; Pl. 3) was done in an 8-day period in August-September 1952. In seeking the tracks previously sketched, the *A* group (Pl. 3) was found and surveyed, perplexing the writer with mild discrepancies, in spacing and azimuths, from prior notes. During mapping of the *B* group, the *N* group was rediscovered (Fig. 2), and the existence of two extremely

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Plate 1.—Stone Tracks and Playa

Figure 1.—Southward View of Stone and Furrow

Track extends 62 feet on azimuth 14° (generalized). Rock measures 10 by 8 by 7 inches, weighs about 60 pounds.

Figure 2.—Northward View of Same Furrow

Scriber rock undoubtedly remained long time in the 1-inch depression before making track. Location at *A* in Figure 1.

Figure 3.—Launching Site of Large Dolomite Blocks

South view of base of steep hill slope bordering playa. Block in foreground measured 11 by 6 by 5 inches, weighs about 30 pounds. Its track is 194 feet long, begins at shore and extends over clay, then across a stony pavement, and again over clay; mapped at *F* in Figure 2.

Figure 4.—Trails Inscribed by Wild Jackass Droppings

Similar signatures of trails, and severe grooving of clay by such light objects, suggest scribers were frozen in floe ice, rather than freely blown by wind. Wild jackasses frequent this area. (L. G. Kirk photo).

Figure 5.—Playa From Northwest

Dark islandlike mass of Grandstand (arrow) is 2 miles distant. Black dolomite hill which has furnished many erratic blocks (Fig. 2) appears above the Grandstand, is 4 miles away and a fraction as high as the mountain behind it.

FIGURE 1

FIGURE 2

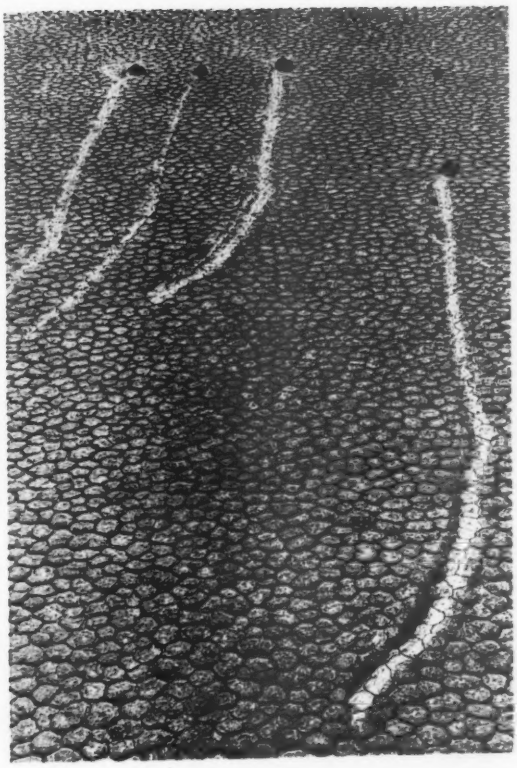

FIGURE 3

FIGURE 4

FIGURE 5

STONE TRACKS AND PLAYA

FIGURE 2

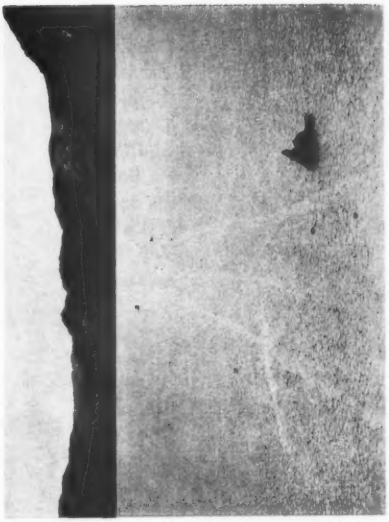

FIGURE 2

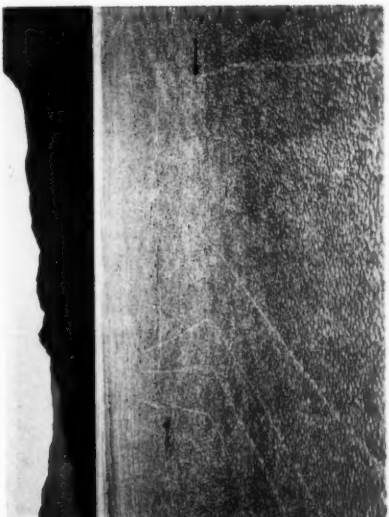

FIGURE 1

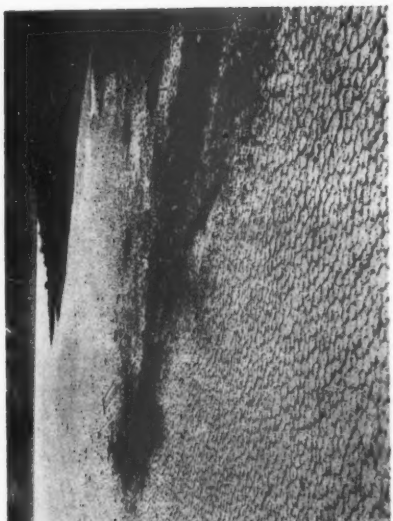

FIGURE 1

GROUPED TRACKS OF SMALL STONES; ICE RAMPART

similar groups was realized. Small trails are easily missed and are viewed plainly only along their courses; from 10 feet aside many are invisible. After the reason for the above differences was understood from office plots, the playa was revisited for 2–3 days late in November 1953 and in mid-April 1954 with some additions to mapping, including newly formed tracks (Fig. 5; *Y* in Fig. 2).

Acknowledgments

Acknowledgment is made for Ray Brodersen's congenial help in principal mapping, for numerous courtesies from George Stough of the Lippincott Lead Mine, for advice from Professor Robert Sherlock of the University of Michigan, Chairman of the American Society of Civil Engineers' Committee on Wind Forces, and for information and co-operation from prospector Joseph H. Crook of Fallon, Nevada; Floyd Keller and Louis Kirk of the National Park Service; Roscoe M. Smith and James F. McAllister of U. S. Geological Survey; James C. Arnold of Gamma Surveys, Inc.; Horace R. Byers of University of Chicago; John Shelton of Pomona College; William S. Housel and James T. Wilson of University of Michigan; Robert F. Bacher of California Institute of Technology; E. L. Felton of U. S. Weather Bureau in San Francisco; Brigham Arnold of University of California; and L. L. Parish of Pacific Telephone and Telegraph Company; also to my son Pat for surveying assistance.

For careful reading and criticism of the manuscript, the co-operation of J. Harlen Bretz and John S. Shelton is much appreciated.

Racetrack Playa and Environment

Racetrack playa, about 3 miles long and 1¼ miles wide, lies at an elevation of 3708 feet (Fig. 1) in a north-south valley east of the Panamint Range, in the northwest part of Death Valley Monument (Pl. 1, fig. 5). The topographic position is best shown on the Ubehebe Peak quadrangle sheet (U. S. Geological Survey, 1951). North-south ranges rise

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Plate 2.—Grouped tracks of small stones; ice rampart

Figure 1.—Southwest view of *A* tracks

The six traces are *A*-1, 5, 6, 7, 2, 8 from right to left; tab 56 of *A*-1 in right foreground (Pl. 3); right arrow points to tab 57. All make common turns near center of photo, evidencing unit movement. Azimuth from distance to turn 49½°, from turn to foreground 70¾°, (from field plots, Pl. 3) altering course by 21¼°. Compare with Figure 2 of this plate. Note offset in *A*-8 at left arrow (Pl. 3).

Figure 2.—Southwest view of *N* tracks

These traces make turns (near tissue box) comparable to *A* tracks, at distance of 207 feet from them (Fig. 2). Azimuth before turn 45°, from turn through foreground 58¼° (by compass in field). Alteration in course only 13¼°, or 8° less than in *A* traces which were made synchronously by same rotating ice floe. Azimuths computed by formula in Figure 3, based on position of *N* tracks relative to *A* tracks, are 44¼° and 58°, close to observed ones. Small track in center foreground, made by cartridge clip, not related to others.

Figure 3.—Northeast view of ice-rampart spit

Formed by ice floes dragging stones obliquely away from shore. Note scraped appearance of its nearer edge, typical of all ramparts mapped. Stones diminish in size toward north or northeast. Note three small stone tracks behind end portion of spit (U.S. Geol. Survey photo).

Figure 4.—Southwest view of *N*, *B*-3 tracks

Viewpoint is 50 feet northeast of that of Figure 2 above, which may be recognized by comparing. The grouped *N* traces, as in diagram (Fig. 2), obliquely transect *B*-3 beneath sack on right.

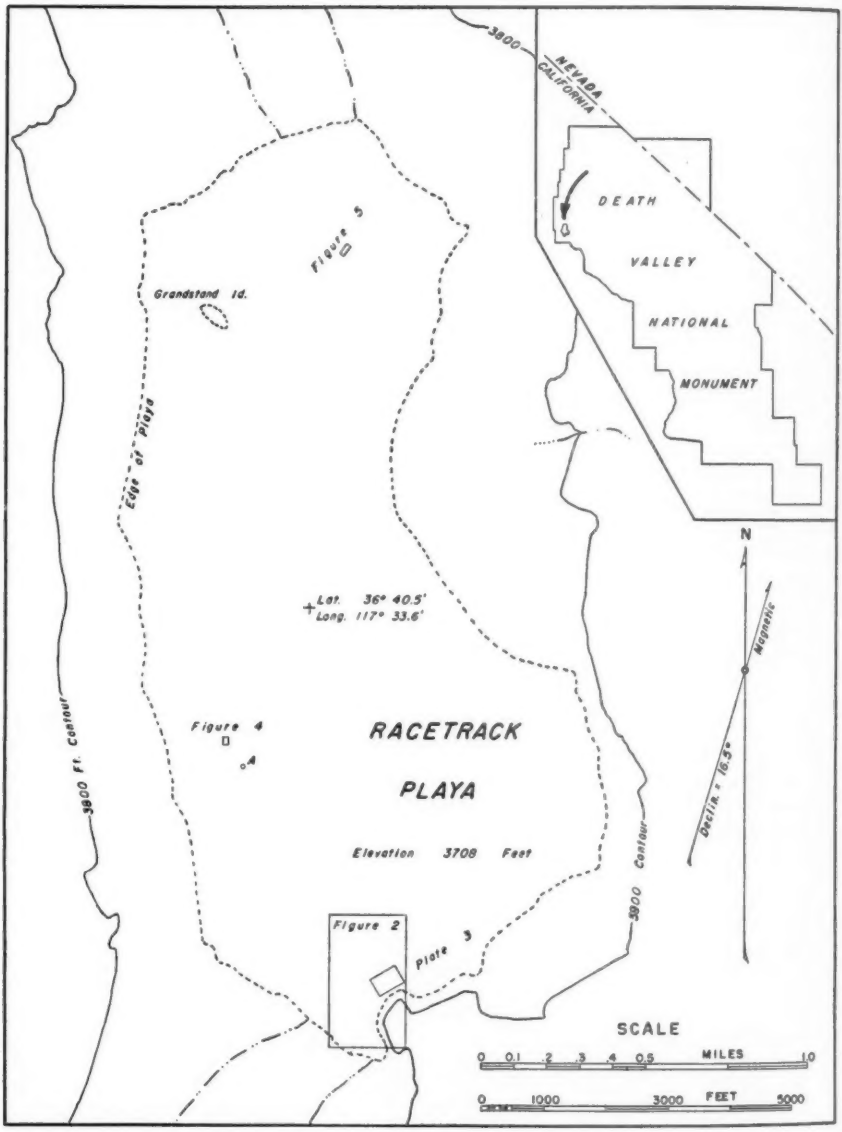


FIGURE 1.—OUTLINE OF RACETRACK PLAYA

General location of playa shown by inset. Positions of Plate 3 and Figures 2, 4, and 5 designated. At *A*, and in north part of Figure 2, are many large, erratic blocks of black dolomite which have made tracks and were derived from promontory in southeast part of Figure 2 where 3800-foot contour indicates steep slope adjacent to shore. (Outline after Ubehebe quadrangle sheet, U.S.G.S.).

from 1500 to 2000 feet above it on both east and west sides. Precipitation is very low, perhaps 3 inches annually. The total drainage area is about 70 square miles, and more than three-fourths of this goes into the north end of the playa. The main tributaries come from the eastern mountains, and they supply large fans both north and south of the playa and a lesser one on the east shore.

The usually dried clay surface of the playa furnishes a hard pavement, smooth and level. Near the south end, the writer's level readings on its surface were uniform and showed natural variations of as much as 0.06 foot only at the indentations of stone tracks. Yet across the narrow, polygonally arranged mud cracks, too light for insertion of a stadia rod, the surface frequently varied more. There is some indication that the south part of the playa surface is about 2 inches lower than the northern part. This is logical since most of the suspended clay must wash in at the north end.

Stones are scattered over the playa. In some places hundreds of feet from shore there were 5–10 small stones (1–3 inches) per square yard but over most of the surface they are less abundant. The frequency of stones diminishes rapidly with increasing size. The only evidence for down-grade transport by washes or mud flows is in a limited area at the north end where the largest wash enters.

Among the stones examined by the writer, three lithologic types predominate on the playa: (1) syenite, frequently subrounded, most abundant on the west side of playa, especially from midway to northern end; (2) dolomite, blue-gray with white bands, also subrounded; and (3) black dolomite, nearly always in angular joint blocks or slivers, the most common type, comprising nearly all large erratic rocks in southern half of playa. Bedrock crops out close to the playa at three places: (1) at the Grandstand—syenite (Pl. 1, fig. 5); (2) off mouth of the eastern wash entering north end (Fig. 1)—low knobs of banded dolomite; and (3) at steep promontory paralleling east shore at south end—black dolomite (Figs. 1, 2). Only at the last can mass wasting deliver large rocks in quantity to the shore. Specimens of all rock types have been observed at the opposite end of the playa from their associated outcrops, but the varying abundance of types (1) and (3) is in harmony with the predominating directions of stone tracks in different quarters of the lake bed.

Grouped Tracks Indicating Unit Movement

General

Fifteen or 20 cases of "parallel" tracks were observed on the Racetrack. Some were merely paired trails, a few feet apart and somewhat like those of a two-wheeled cart, except at bends the turns were not usually opposite one another and spacing of tracks changed (Pl. 3). Two carefully mapped groups demonstrate the relationships.

Method of Mapping

For making the detailed plots (Figs. 4, 5; Pl. 3) numbered tabs of tracing cloth were nailed to the clay along the trails selected. Nails were driven vertically, but not home, allowing measurements to shank. Base lines were then aligned by alidade alongside the trails, with stations at intervals of 25 or 50 feet marked by tabbed and lettered chaining pins. Distances of each tabbed point on the tracks, from two or more base line stations, were measured with a 100-foot steel surveyor's tape, read to nearest 0.005 foot. The three base lines of Plate 3 were interrelated by eight measurements, from each of the two stations II-*N* and II-*P*, to I-*A*, I-*C*, IV-*A* and IV-*C*. In later office work, and as a safeguard for concluding shift of the *C* scriber, trigonometric computation of triangles in net formed by these lines was carried out, with discrepancies all less than 0.05 feet.

After base lines and tabbed points had been laid out on plane table (1 inch = 10 feet), tracks were sketched between tabs while occupying progressive viewpoints along them. There are undoubtedly faults in this sketching, owing largely to unavoidable perspective and oblique view which highly accentuates bends and irregularities. Partly for this reason the minor wiggles of *A* tracks, undertaken first, appear more exaggerated than analogous parts of *B* or *C*. Anticipation of inaccuracies in sketching

prompted greater care in horizontal control of tabbed points. All measurements were replotted in the office with a fine needle point and field sketches of tracks were transferred to the new map.

The south end of the playa (Fig. 2) was mapped by plane table, including tracks *G*, *H*, *L*, *J*, *D*, *E*, and *F* with numerous stadia-rod positions for last three; tracks *P*, *Q*, and *R* only by compass azimuths and pacing in conjunction with photos and field sketches, and their locations by stadia positions for the large rocks; track *Y*, not yet formed at time of mapping, was added by later field sketch, and those at *S*, already rather obscured, by McAllister's prior photo (about 1947–1948; Pl. 2, fig. 3). The *A*, *B*, and *C* tracks were added to Figure 2 from original measurements (as in Pl. 3), but those at *M*, *N*, *T*, *U*, *V*, *and X* in the following manner. After office study showed mathematical agreement among *A*, *B*, and *C* trails, and with knowledge that other similar tracks (*N*) were present near by, the writer located by plane table three critical points comparable to tabs 50, 57, and 58 of *A*-1, for each similar track that could be found. Compass azimuths and taped distances of legs between these critical points were also measured for each track. Both overlay-test and trigonometric calculation then proved these values and the identity in signature with *A*, *B*, and *C*; except only that *X*-1, 2 were fragmentary and furnished only one critical point each. The tracks plotted through these critical points, with slight extensions beyond them conforming to photos, field sketches, and in a few cases memory, lack the precision of those in Plate 3.

Signature Group Near South End of Playa

General description.—These tracks are 1–2 inches wide and 200–300 feet long (Fig. 2; Pl. 2, figs. 1, 2, 4; Pl. 3). No single photograph from the ground can show more than a portion of them. The four longest *A* tracks (*A*-1, 2, 3, 4) were tabbed in detail, and four more (*A*-5, 6, 7, 8) were sketched by eye. The striking parallelism of seven of them, excluding *A*-4, had drawn attention. The similarity of *A*-4 and *C* to these seven, though in more offset positions, led to tabbing and plotting them also. As anticipated in the field, in the office they were found to have been made by the same unit movement as the other *A* tracks. The three *B* tracks were better disguised. They bear little resemblance to the others (Pl. 3). Though tabbed and plotted for their mutual similarity, relationship to the *A* and *C* tracks was unsuspected until proved by sliding overlay tests in the office.

Differences between these 12 inscriptions are obvious. For instance, *A*-4 between tabs 83 and 69 has less azimuth than corresponding stretches of the other *A* and *C* tracks; it bears closer to north. This is also true of *B*-1, 2, and 3, northeast from tabs 82, 20, and 51 respectively, and increasingly so in this order.

Corresponding parts of *C* and *A*-1 are unequal in length; *C* is 166 feet from the turn near tab 14 to tab 34, whereas it is only 114 feet from tabs 57 to 45 along *A*-1. Corresponding lengths of *B*-1, 2, 3 are likewise longer, part for part, than *A* tracks, and increasingly so in this order, but all shorter than *C*. Summarizing, no tracks in Plate 3 are strictly identical or parallel, but differences diminish with proximity. All represent the same signature. Completeness of the signature varies in different tracks and shows most fully in *A*-2, next so in *A*-1. The *B* markings are less complete and lack the curving, northeast part of the signature, which, plus the considerable difference in their courses, delayed recognition of signature identity with the others.

The map of these trails does not indicate quite their total extent. Most became indistinct gradually a few yards beyond tabbed limits. The southwestern ends of *B*-2 and *B*-3 were clear and were tabbed, however, and the northeastern termini of *B*-2 and *A*-5 were mapped on a later visit.

Disparities of length and course are resolved by testing the mapped tracks with a transparent overlay. Analogous points may be chosen and plotted on the overlay, one for each track to indicate stone position, and the transparency may then be moved to ascertain whether stone positions continue to follow proper tracks. This experiment is crucial in proving ice-floe origin. To maintain registry of points over tracks, the overlay must rotate as it progresses except near northeast ends and altogether 18½°. Rotation caused the disparities.

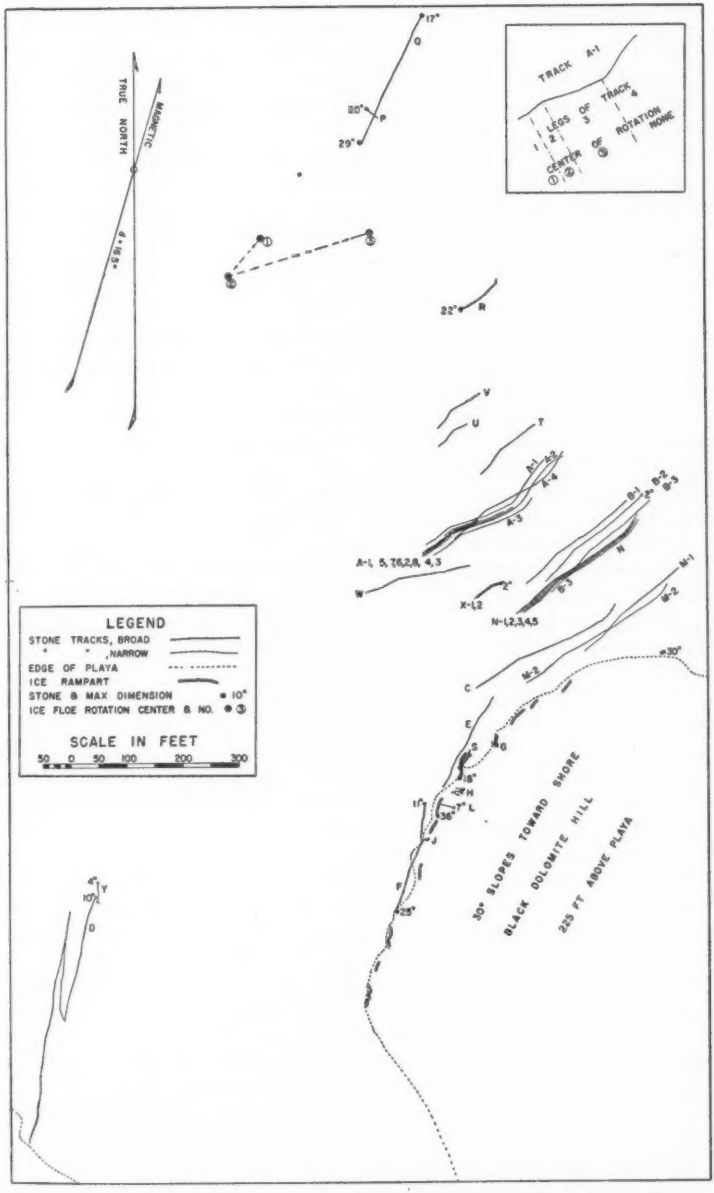


FIGURE 2.—MAP OF SOUTH END OF PLAYA SHOWING STONE TRACKS AND SHORE

Tracks *Q*, made by 230-pound rock at north end, and *R* by 320-pound rock at west end, are broadest and most conspicuous. Longest track on playa is *D* (872 feet). When it was mapped (September 1952), track *Y* was nonexistent. All 25 tracks between *V* and *M*-2 have identical signatures, were made by small stones frozen to thin ice flow of insufficient strength to move large rocks. Signature is divisible into typical legs, as in upper right inset. Obstructed by large rocks (in Track *R*, etc.) and scraping shore, ice flow made legs 1, 2, and 3 by counterclockwise rotation on indicated centers. Meanwhile rock in *R* was slicing through floe. At end of legs 3, floe split on this work, and the section of ice over tracks continued northeast, made legs 4 without obstruction or rotation. Cusps and wiggles common in earlier parts of tracks are absent from legs 4 (Plate 3).

Overlay tests demonstrate that all 12 trails in Plate 3 were made by unit movement of the scribers which were necessarily linked by planar body at least 200 by 275 feet. Only a sheet of lake ice could fill this role. Further evidence proves it was even larger though it broke in two before completing inscriptions. Additional tracks of the same signature (Fig. 2), together with circumstances causing rotation, prove that the floe was about 1000 by 300 feet before breaking.

Rotation was not anticipated when the writer undertook mapping of tracks which appeared identical. Although mapping proves unit movement as foreseen, it also proves that no tracks are identical, and that all 25 stones which inscribed this signature assemblage (Fig. 2, *V* to *M*-2) were traveling with different speeds and directions.

Direction of travel.—The lack of stones in termini of the *A*, *B*, and *C* tracks delayed definitive conclusion on direction of movement. Perhaps 10 per cent of the stone tracks on the playa have no stone at either end. This may be due to indurable objects (Pl. 1, fig. 4) taking the role of stones (Shelton, 1953, p. 438) or to rafting away of stones as water depth increased by change of wind.

Close inspection of transections of track *A*-4 with the other *A* tracks indicated that *A*-4 was engraved prior to *A*-1 at tab 85 and near tab 73; also *A*-4 was prior to *A*-2 at tabs 88 and 71. This relationship was definite at tabs 85 and 88, but less so at 71 and 73, and no inference could be drawn from the other transections at tabs 62 and 89. Also *N*-3, 4 cut *B*-3 on enlarged photo (Pl. 2, fig. 4). These indicated movement northeastward, like predominant movements over this part of the playa.

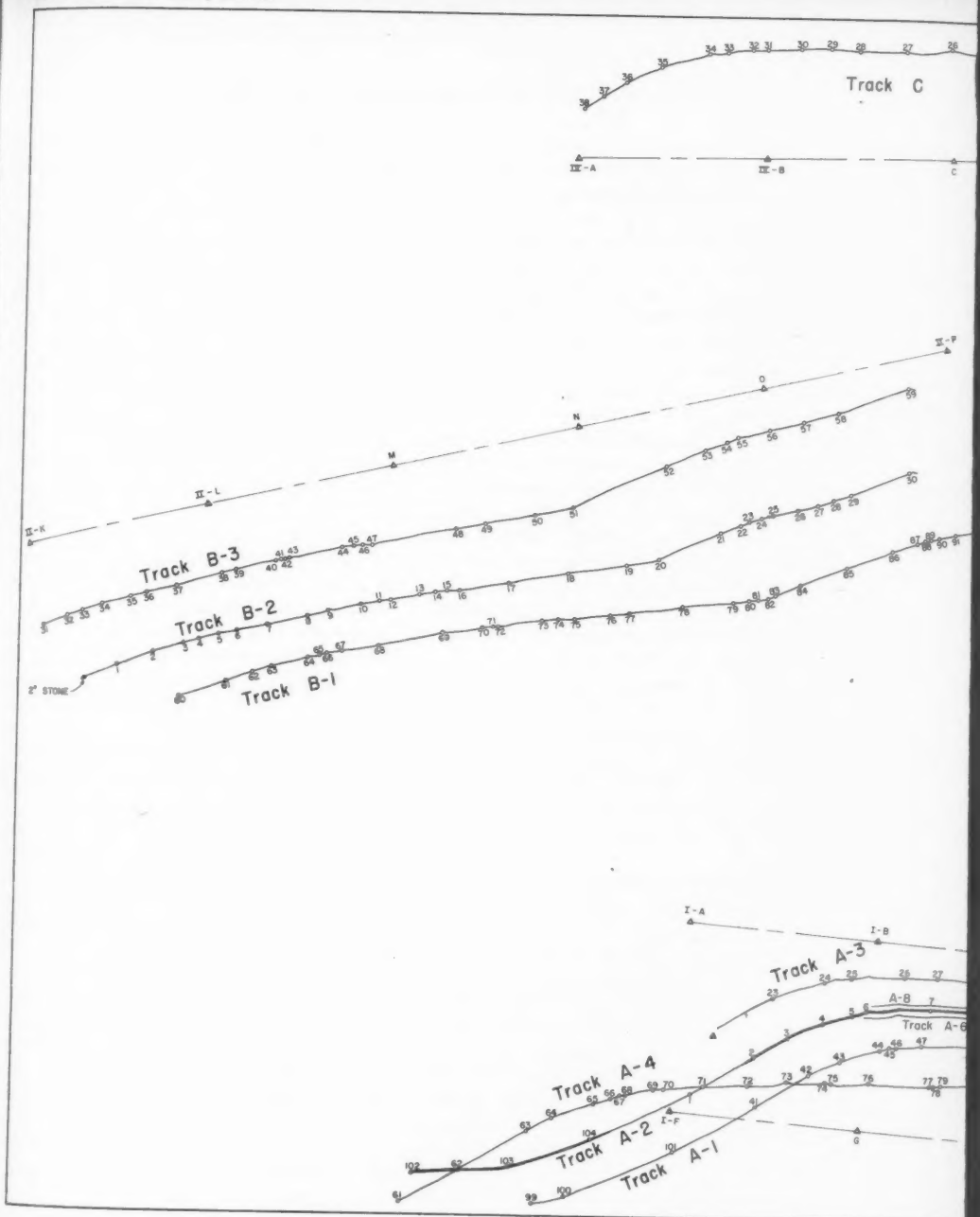
Later visit to the playa confirmed this. A triangular prism of dolomite, 2½ by 2 by ¾ inches, was found in the faint northeast end of the *B*-2 track almost 10 feet northeast of tab 1. The *A*-5 track also extended 9 feet farther northeast than previously mapped, with a stone, 1¾ by ¾ by ½ inches, at its terminus (Pl. 3). A stone occurs also in northeast end of *X*-2 (Fig. 2).

Length of trails.—The longest tracks are *A*-2, 319 feet; *C*, 295.7 feet; and *A*-1, 275.3 feet. In completeness *A*-2 would come first again, but *A*-1 would surpass the longer *C* track, and so would *A*-3 which is 213 feet long. The northeast part of the signature occurs only in *A*-2 (Pl. 3), but only the sliding overlay shows that the southwest extremity is written only in *B*-2, *B*-3, and perhaps in *A*-4. The mathematical relationship between lengths of tracks is shown in Figure 3 and details of quantitative comparisons in Table 3. For instance, 145.60 feet of *B*-3 (less than its total length) was synchronous and comparable to 110.35 feet in *A*-1.

Rotation of ice floe.—The total rotation of the ice floe was 18½° counterclockwise in traveling 188.5 feet along path of track *A*-1 (tabs 60-45). Figures for rotation during short segments of travel along *A*-1, progressing from southwest to northeast, are listed in Table 1. These were measured by protractor from the sliding overlay. Rotation averaged 9.6° per 100 feet of travel (Table 1, Col. V), from tabs 58 to 57, increased to 10.6° from tabs 57 to 45, and there abruptly ceased. The significance of these facts is brought out below.

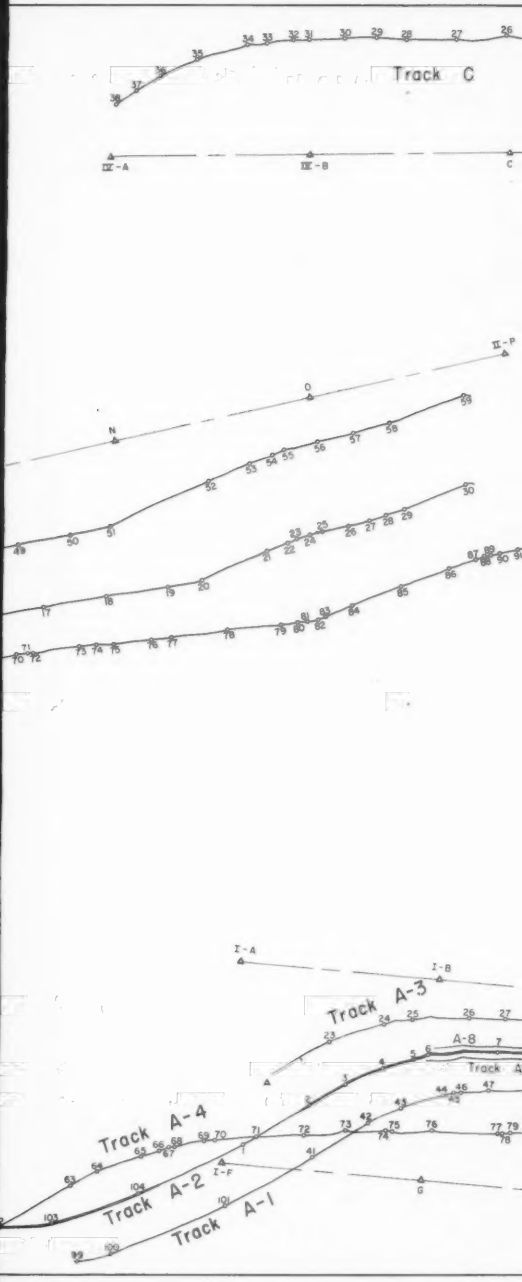
The rotation could not have been caused by the ice floe brushing the shore, 60 feet southeast of track *C*, since the retarding effect would have applied opposite torque. Yet it surely did rub against shore.

The cause of rotation was not clear until centers were determined. Graphical methods proved better than calculations. This required a transparent overlay 9 by 3 feet, in a light, rigid frame. One end was moved along Plate 3 (1 inch = 10 feet) to match tracks in various positions, while center was observed at the other end by identical grids, one stationary, the other on overlay. Center was observed after movement, not during it, and for various amounts of movement: (1) with these in some cases adjusted to span between prominent wiggles in tracks, (2) for all successive tab intervals of *A*-1, and (3) for principal legs of tracks as diagrammed (inset, Fig. 2), using intervals between tabs 60, 58, 57, and 45 of *A*-1. The last are most significant and are plotted (Fig. 2). Experimental arcs drawn on these three centers (but not illustrated) very closely match corresponding legs of all tracks, and chord lengths (for 23 tracks) are closely proportional to respective radii. The arcs superimpose perfectly on legs 2; they deviate a little from



MAP OF TWELVE STONE TRACKS WITH I

Tabs nailed to tracks by numbered circles, surveyed by two or mo to one of stations, triangles, at 50-foot intervals along base line made tracks are indicated by tabs 57 of A-1, 83 of A-4, 82 of B-1, e overlay, and sliding it with proper rotation, marked points follow is at zig-zag of B-2, tabs 24-25, explained in Fig. 3. Cusps and wi ice floe brushing irregularities of shore line (Fig. 2) and are ab tab 45 of A-1, where ice floe was freed from obstacles and ceased this plate may involve discrepancies due to optical distortion.)



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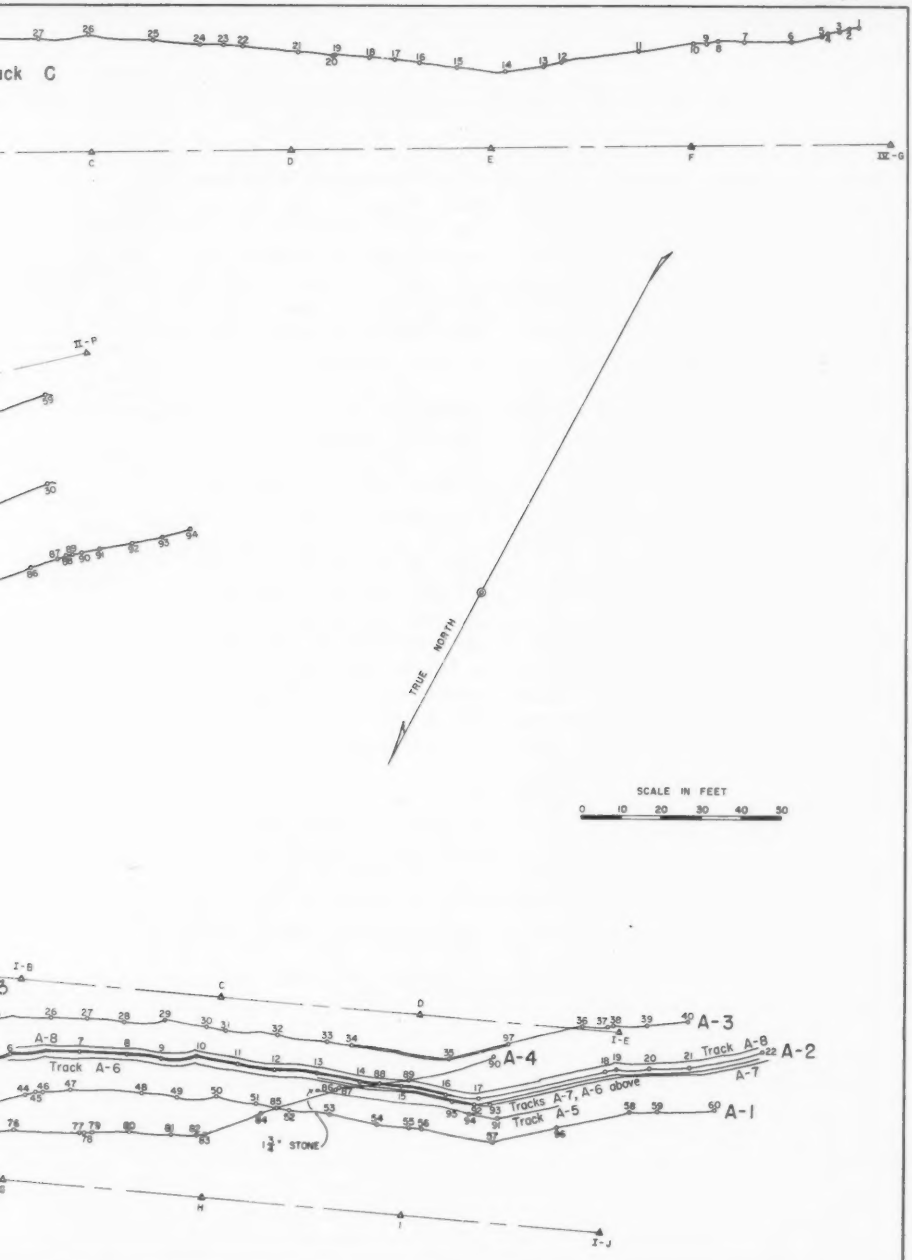
d sliding it with proper rotation, marked points follov
g of B-2, tabs 24-25, explained in Fig. 3. Cusps and
ushing irregularities of shore line (Fig. 2) and are
-1, where ice floe was freed from obstacles and ceas
ay involve discrepancies due to optical distortion.)



WITH IDENTICAL SIGNATURES

by two or more taped measurements from each tab ong base lines. Relative positions of stones which , 82 of B-1, etc. By marking these on a transparent oints follow respective tracks. A notable exception Cusps and wiggles, as at tab 50 of A-1, were due to and are absent from legs 4, northeastward from s and ceased rotating. (Overlay tests made from stortion.)

legs 3 (maximum 0.15 inch = 1½ feet) but *are always on the same side of analogous parts of all tracks*; of course they do not match minor wiggles of tracks. The essentially arcuate character of legs 3 may be seen by looking obliquely along them (tabs 20-1 of *B*-2, or 57-45 of *A*-1; Pl. 3). Legs 1 are short, and southwest ends not well delimited. The center of rotation made two major shifts between positions 1-2-3 as illustrated (Fig. 2). Methods (1) and (2) of locating center, above, show also, as substantiated by positions of arcs with respect to tracks, that center was not fixed but shifted slightly about each of the three positions, excepting perhaps position 2. The centers illustrated are merely mean centers for their respective legs. These demonstrated complications fit theoretical considerations below. Summarizing, the center of rotation of the ice floe was always in the vicinity of large rocks far northwest of the tracks (Fig. 2), never closer than 100 feet to any of them, and was mobile with two principal changes, the larger one toward the 22-inch rock in track *R*. Encumbrance by large rocks caused the rotation.

The great force imparted by strong wind to an extensive floe would exceed the strength of thin ice bearing on obstacles like the large rocks referred to, and contacting ice would crumple. Crumpling of a rotating floe against one pivot rock would displace the effective center of rotation away from it. For the case in hand, while the floe rotated 12¼° making leg 3 of *A*-1, if ice was forced *northeast* against rock in *R*, using it as pivot, failure of 1 foot of ice here would displace the effective center of rotation 5 feet *northwest* of rock; if ice broke when forced *north* against rock, center would shift *west*; if forced *east*, would shift *north*, etc. A *southeast* shift of center would be caused by floe wedging between rocks and shore, and the shore squeezing it slightly in *northwest* direction. Because of crumpling, several scattered rocks might act jointly as pivot.

The remoteness from rocks and minor shifts of centers are explained by the preceding. The small wiggles and cusps of tracks are probably due to slight movements, northwest and southeast, of the rotating floe while scraping against ice ramparts and changing shore configuration. The largest shift in center, from positions 2-3, probably represents the moment when the forward edge of the ice reached rock in *R*, and this replaced more distant rocks as pivot. Immediate crunching of ice there would have been likely. At this moment the major bend in tracks was inscribed (tab 57 of *A*-1; Pl. 3). Scrutiny of enlarged photos (Pl. 2, figs. 1, 2) shows that tracks approach this turning point straight; northeast of it for 5 feet they curve slightly, then gradually straighten. This delicate curve was noted when tabs were first set and prompted nailing 82 of *A*-4, and 80-81 of *B*-1. Though unapparent in plan (Pl. 3), it was plain in oblique view. The curves resemble the easement curves adjoining tangents of railway tracks, and were caused by crunching of ice. After this turn the tracks indicate increased angular velocity of ice floe, as shown by Table 1 and in proportion to decreased radius from new center.

During inscription of legs 3 of the tracks, the ice floe was continually cut by rock in *R* for a total of 45 feet, which finally split the floe into two parts. The part containing the inscribing stones was then freed, thereafter moved northeast unimpeded by rocks or shore, and changed direction only as wind varied. The split occurred at tab 45 of *A*-1, and northeast from there the tracks are smooth curves without cusps or wiggles and prove by overlay that the floe had ceased rotating.

Analogous points for sliding overlay test.—The test for unison movement requires selection of analogous points on all comparable tracks to show relative stone positions. The points are critical for mathematical treatment. For trails in Plate 3 the most obvious group of points is Set I in Table 2. After marking these on an overlay, and moving it to new positions, the other sets may be easily found. Discrepancies between sets probably stem from too casual placement of original tabs and from errors in mapping or plotting the tracks, but in some cases they demonstrate shifted positions of stones in the ice floe. With the eight points of Set I a surprising registry of points over trails is found by moving the overlay. In performing overlay test, one should remember that the *B*-2, *C*, and *A*-3 scribers shifted, and that in between tabbed positions the tracks were plotted merely from sketching. Among the 490 measurements taped to determine tab positions (Pl. 3) there

may have been errors; measurement of distance from a third base-line station, as a check on location, was applied to only 24 of the 233 tabs.

the *B*-1 point unvarying, scatter for all other points is obtained (Table 2, Col. VIII). For only one of these points (*A*-4) is there appreciable

TABLE 1.—ROTATION OF ICE FLOE DURING TRAVEL

I Position of point 57 on track *A*-1	II Cumulative rotation (*in degrees*)	III Rotation along leg (*in degrees*)	IV Length of leg in *A*-1 track (*in feet*)	V Rotation rate (*in degrees per 100 feet travel*)
*	−6.25			
		1.0	16.8	5.94
Tab 60	−5.25			
		1.25	14.3	8.74
Tab 59	−4.0			
		0.625	7.0	8.84
Tab 58	−3.375			
		1.125	11.6	9.7
7 ft. SW. of tab 96	−2.25			
		0.75	7.0	10.7
Tab 96	−1.50			
		1.5	16.4	9.15
Tab 57†	0			
		2.0	18.1	11.05
Tab 56	2.0			
		1.25	11.2	11.16
Tab 54	3.25			
		1.50	12.0	12.50
Tab 53	4.75			
		0.75	10.2	7.30
Tab 52	5.50			
		1.0	8.4	11.9
Tab 51	6.50			
		1.2	10.1	11.89
Tab 50	7.70			
		0.8	9.9	8.08
Tab 49	8.50			
		1.0	8.9	11.22
Tab 48	9.50			
		1.75	17.8	9.83
Tab 47	11.25			
		1.0	8.8	11.35
Tab 45	12.25			
		0	13.7	0
Tab 43	12.25			
		0‡	9.0	0
Tab 42	12.25			
		0‡	16.2	0
Tab 41	12.25			
		0‡	24.9	0
Tab 101	12.25			
Total	18.50	18.50	252.3	

* This position beyond end of *A*-1 track; position defined by *B*-2 and *B*-3 scribers at tabs 30 and 59 respectively

† Zero position arbitrarily set at most critical point in track

‡ Almost negligible, about ¼° clockwise, in reversed direction, tabs 43–101

Choice of significant positions for nailing the tabs allowed leeway frequently of several inches along the track; it was little transversely because of slenderness of the trace. Accordingly, when all sets of analogous points in Table 2 are plotted, the spread for any one is almost entirely longitudinal.

Evidence for shift of the *C* and *A*-3 scribers necessitates plotting all points in Table 2. Any one point of the first set, when plotted, might be assumed unvarying and centered over a tab, while other sets are marked down. Similar results are obtained whether the unvaried point belongs to *A*-1, *B*-1, or *B*-3; but use of *B*-2, *C*, or *A*-3 points in this way would be misleading because of shifts, and *A*-2 is suspect. Assuming

spread transverse to the track; this amounts to 0.50 foot and is probably due to one incorrectly measured distance.

Labelling points on overlay with numbers of tabs that indicate them reveals that displacements of *C* point are all in a lagging direction. This indicates scriber shift. Similar behavior is evidenced for *A*-3.

Scriber shifts.—A lateral shift of a couple of inches may be seen in track *A*-8 about midway between tabs 17-18 on the neighboring trace (Pl. 2, fig. 1; Pl. 3). Probably the stone overturned or spun around in such a way that it, or an inscribing corner of it, was offset from the earlier location.

Most shifts were longitudinal, and the largest

TABLE 2.—IDENTICAL POINTS OF SIGNATURE IN *A*, *B*, AND *C* TRACKS

Track	Sets of analogous tab points							Scatter (*in feet*)‖
	I	II	III	IV	V	VI	VII	VIII
A-1	57	50	58	59	45	51	nt¶	0.7
A-2	17	10	19	20	6	11	....	1.4**
A-3	35	29	36	38	§	30	nt¶	3.0**
A-4	83	73	87	§	67	74	....	0.8
B-1	82	69	88	91	§	71	nt¶	0
B-2	20	9	†	25‡	§	11	30	21.8**
B-3	51	39	55	56	nt¶	43	59	0.3
C	14.2*	26	10	8	34	25	nt¶	8.8**

* At angle, about two-tenths of distance, 14 to 15
† *B*-2 scriber idle during inscription of this point in all other tracks
‡ *B*-2 scriber southwest of where it was displaced from floe
§ No tab placed here
¶ nt—Track does not extend this far
‖ Scatter is with position of *B*-1 scriber assumed unvarying
** Scatter due to scriber shift, *A*-2 questionable

TABLE 3.—COMPARISON OF OBSERVED *A* AND *B* TRACKS WITH COMPUTED ONES

Row	Tracks	Tabs* *P*–*P′* *S*–*S′*	Measured quantities‡							Computed†	
			I D_p	II *M*	III *B*	IV *Q*	V *Y*	VI *A*	VII D_s	VIII *A*	IX D_s
1	*B*-3 with *B*-1	56–39 91–69	145.60	56.90	43°30′	11°42′§	37°39′	3°50.3′¶	138.60	3°47.6′	138.84
2	*B*-3 with *B*-2	51–31 20–‖	144.20	26.50	43°00′	11°20′	37°20′	1°36.3′¶	140.70	1°41.4′	141.09
3	*B*-3 with *B*-2	59–56 30–25	38.43	22.10	116°15′	2°15′§	115°07½′	−0°29.3′¶	37.72	−0°33.6′	37.65
4	*B*-3 with *A*-1	56–39 59–50	145.60	252.80	56°30′	11°42′§	50°39′	17°15′	110.35	17°10.3′	110.68
5	*A*-1 with *A*-4	58–45 87–67	148.60	72.80	183°00′	15°37½′§	175°11′	−7°45′	147.83	−7°38.7′	148.26

* Each vertical pair of tabs, *P* and *S*, or *P′* and *S′*, analogous points of different tracks
† Computed from formulae stated in Fig. 3
‡ Determined by measurements from Pl. 3 except as noted otherwise. Distances in feet. Symbols refer to diagram and formulae in Fig. 3 inset
¶ From measurements of tangents in Fig. 3; all other angles measured by protractor
‖ Halfway between tabs 1 to 2 which by overlay proves synchronous with tab 31 in *B*-3
§ From Table 1

and most obvious displacement of inscribing tool was in track *B*-2. Between tabs 24 and 25 is a sharp zig-zag, scarcely 1 foot long, which was puzzling in the field since there was no counterpart in the accompanying traces *B*-1 and *B*-3. The explanation was found, however, after a few trials with the transparent overlay. The *B*-2 stone became detached at the zig-zag, while the ice floe advanced 22 feet without it, and then it was re-attached and resumed its inscription. The following clues to this may be seen in Plate 3: track *B*-2, tabs 20–30, is 24 feet shorter than *B*-3, tabs 51–59, yet the extreme parts of both stretches are respectively comparable. This is 10 times the difference that would result from rotation (Table 3). Another clue is in points at tab 55 of *B*-3 and tab 88 of *B*-1, which are analogous with one another and

with all points of set III in Table 2. This point of the signature, and about 10 feet of track both ways from it, are missing from *B*-2 because of the idle stone. The phenomenon is apparent in Figure 3 which shows actual and adjusted positions of the extreme portion of *B*-2, and the gap left by the missing part of track. The reorientations of trails involved in producing this diagram stimulated a study of exact quantitative relationships.

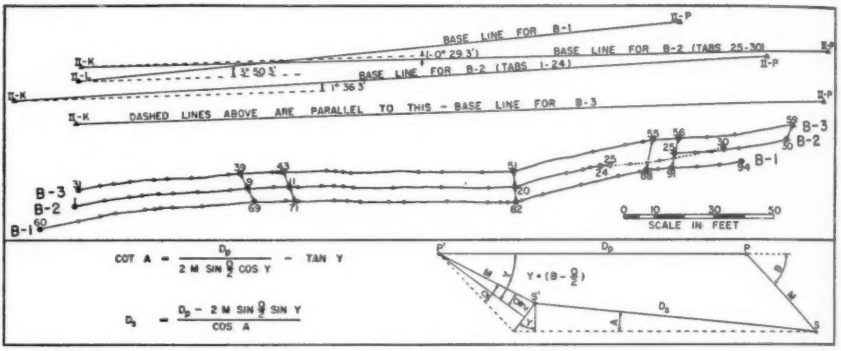

BASE LINE FOR B-1
BASE LINE FOR B-2 (TABS 25-30)
BASE LINE FOR B-2 (TABS 1-24)
DASHED LINES ABOVE ARE PARALLEL TO THIS - BASE LINE FOR B-3
B-3
B-2
B-1
SCALE IN FEET
COT A = Dp / 2 M SIN Q/2 COS Y - TAN Y
Ds = Dp - 2 M SIN Q/2 SIN Y / COS A

FIGURE 3.—REARRANGEMENT OF B TRACKS FOR CLOSE COMPARISON

Arrows indicate analogous points and progressive expansion of signatures, from *B*-1 to *B*-3, away from rotation center. Repetitions of base line at top, in true relative position to each track, show angles of reorienting *B*-1 and both parts of *B*-2 to attain parallelism. Trigonometric computations (from inset formulae) prove these angles and show that inscribing stones were rigidly joined in rotating ice floe. Right end of *B*-2 is moved 22 feet to right from actual (dashed) position to match its neighbors, signifying scriber shift at zig-zag near tab 25.

The *C* stone fell behind four times in its journey. Although two lapses amount to barely 1 foot each, within range of errors, the larger ones are real, and all four are in the same direction. The cumulative retardation was 9 feet. Since the *C* stone moved faster in the rotating ice than any *A* or *B* stones, therefore with greater frictional resistance, its repeated relapses are the more reasonable. Similarly leg 3 of the *M*-2 track is 17 feet shorter than would fit mathematical relations, and legs 2 of *M*-1, 2 both a trifle so.

The *A*-3 stone was dislodged and retarded somewhere in the stretch, tabs 97-34, and this is exactly the part in which its trace is distinctly double. The *A*-2 stone made double traces when traveling more easterly courses, a single one on more northerly ones, between which it probably turned in the ice, but the overlay shows no significant retardation (Table 2).

Mathematical relationships.—Stones fixed at some distance apart would make traces different in length and direction during rotation of the floe. The observed differences agree with theoretical ones computed from trigonometric formulae.

On the inset diagram of Figure 3, two points *P* and *S* are separated by a fixed distance, *M*, in an ice floe assumed to travel to the left, rotating counterclockwise by angle *Q*. The scribers then have reached positions *P'* and *S'* with net travel distances of D_p and D_s, though the route may have been circuitous. Angle *B* is measured clockwise from back extension of line *PP'* to *PS*. The net course of *S* deviates from that of *P* by angle *A*, also measured clockwise. Formulae are stated for the angle *A* and the distance D_s.

Computations by these formulae were applied to corresponding stretches of tracks *B*-3 and *A*-1. As measured from the plot of Plate 3, the net travel of the *B*-3 scriber was 145.60 feet and of the *A*-1 scriber 110.35 feet. The computed travel of the latter is 110.68 feet. The protractor-measured divergence in courses is 17° 15′, the computed one 17° 10.3′ (Table 3).

In Figure 3 trails *B*-1, *B*-2, and *B*-3 have been taken from their true positions in Plate 3 and rearranged so as to place analogous points (indicated by arrows) side by side. The trails are reproduced much closer together, and also bodily reoriented to attain optimum parallelism

by eye judgment. The base line is repeated in relative position to each trail, and the angles of changed orientation for *B*-1 and *B*-2 with reference to *B*-3 are shown between base lines. Shown this way the tracks are more readily compared. The right end of *B*-2 track, tabs 25-30, was disconnected at the zig-zag, beyond which overlay shows clashing disagreements, and moved right by the determined amount of scriber shift; and it is reoriented independently from the rest of *B*-2. Without severing, shifting, and reorienting it, this fragment of track is discordant with adjacent ones, for it was made earlier than portions of *B*-1 and *B*-3 beside its dashed position.

Computations for the relations of *B*-1 to *B*-3, and for both the left and right parts of *B*-2 to *B*-3, are shown in Table 3. The agreement between observed and computed lengths and deviations of courses is good. That eye judgment on parallelism achieved within 6 minutes of arc from a theoretically correct angle surprised the writer in all three cases. It is entirely logical that the *B*-1 track had to be reoriented through a greater angle than *B*-2 in order to effect visual parallelism; its natural position is farther from *B*-3; and that the shifted end of *B*-2 had to be swung in opposite sense, for the lead had passed from one track to the other because of the shifted stone.

Computations uphold the interpretation that the tracks were made by unified movement rather than by independently moved stones. No calculations were made for tracks *A*-2, *A*-3, and *C*, because scriber shifts at unspecifiable points disallow similar mathematical treatment.

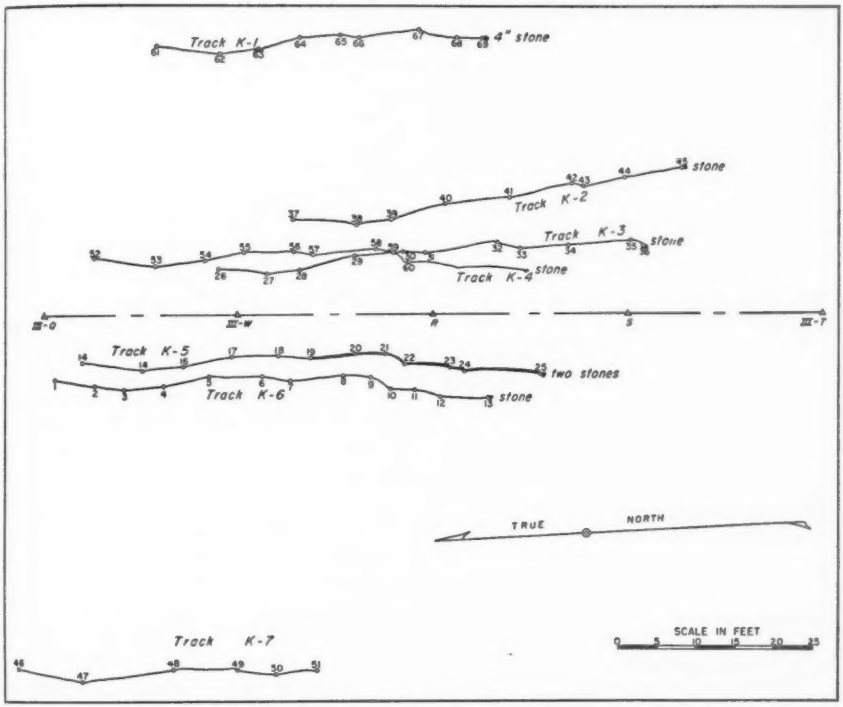


FIGURE 4.—MAP OF SEVEN STONE TRACKS WITH SIMILAR SIGNATURE

Positions of tabs on tracks shown by numbered circles, measuring stations at 25-foot spacing on base line, by triangles. All signatures identical in northern parts. Ice floe rotated counterclockwise, moving stones south, but broke into two parts at tabs 67 of *K*-1, and 21 of *K*-5, after which *K*-1, 2, and 3 continue one signature, *K*-4, 5, 6 a different one.

Signature Group Near West Side of Playa

Ten tracks were mapped here, but only seven are illustrated in Figure 4. The three easterly ones mapped by plane-table and stadia readings

TABLE 4.—COMPARISON OF OBSERVED K TRACKS WITH COMPUTED ONES

Track	Tabs S-S′	Measured quantities* I M	II B	III D_s	Computed IV D_s
K-0	†	155.50	89°40′	11.38	11.40
K-1	62–66	82.675	104°25′	17.885	17.97
K-2	38–41	68.275	122°55′	19.89	20.00
K-3	27–31	57.50	116°16′	20.42	20.41
K-4	53–57	54.225	101°57′	20.22	20.25
K-5	15–19	40.65	102°52′	21.55	21.395
K-6	3–7	37.85	100°05′	21.36	21.60
	P–P′			D_p	
K-7	47–50	...	.....	24.78	...

* M and D_s given in feet. Rotation (Q) was 4°52′ counterclockwise

† K-0 most easterly track, not shown in Fig. 4

D_p, from which values of D_s were computed, is 24.78 feet (K-7)

D_s in column III for comparison with computed D_s in column IV

proved insufficiently precise for comparison with the others. Stones were observed at the south ends of most of the trails.

Overlay test proves that the seven tracks, K-1 to K-7, were made by a southward unit movement with rotation counterclockwise, throughout the earlier and greater part of their length. When the K-6 stone was near tab 9, the ice floe broke apart, as indicated by the differing characters of the two subgroups, K-1, 2, and 3 and K-4, 5, and 6. Trail K-7 ends too shortly to be affiliated with either.

The trails range in width from 1½ to 2½ inches, and the longest (K-5) extends 60 feet. From overlay, two sets of analogous points (S-S′, Table 4) show rotation of 4° 52′ from one position to the other. The intervening stretches have been calculated from the formulae of Figure 3 and may be compared with corresponding scaled distances in Columns III and IV of Table 4.

Evidence of unified movement is strengthened by the westward increasing value of D_s, in trigonometric proportion to spacing of the tracks, from K-1 to K-7. In the most easterly track (K-0, not illustrated), D_s is less than half that in K-7. Rotation caused one stone to travel with more than twice the speed of another in the same floe. Another symptom is in relative trends of legs in the last tabbed stretch before the S′ position. In K-7 (tabs 49–50) and in corresponding stretches of K-5, K-4, and K-1, travel was a little west of south. In K-3 the scriber preceded all these and was being swung relatively eastward by counterclockwise rotation. Its course (30-31) is almost due south. In K-2, with still more scriber lead and eastward component, the course (40-41) was east of south.

Splitting of the ice floe is indicated by the two contrasting types of turn to the southwest. It is a sharp turn in the eastern subgroup of tracks, as at tabs 42-43 in K-2, and a curving turn in the western subgroup, as at tabs 8-10 in K-6. Moreover, the two categories clearly depart from one another beyond here. The western section experienced additional rotation of about 9°; the eastern section about 13°, or a total of about 18° from its orientation at the north end of the tracks.

If stones are blown independently by the wind some might move farther or faster than others. To interpret these stones as independently blown would hardly match the fact that the stone in each track, from east to west, was moved progressively faster, in trigonometric proportion to spacing, and in complete disregard to fortuitous disparities in shapes and sizes. Similar argument could be applied for the differences in courses of the tracks.

OTHER TRACKS

Additional tracks argue against origin by freely blown rocks. The longest stone track recorded is detailed at D in the map (Fig. 2) and in Table 5. It has three legs, mildly deviating from straight lines, and is 872 feet long and 8 inches wide. An elongate, 27-pound, dolomite boulder is at the terminus, and its long axis was transverse to the track during the entire inscription.

The boulder is so nearly round it is impossible to suppose that winds would move it without

rolling it. The trail is positive evidence that motion was a sliding one, which would be natural for a stone in an ice floe.

Stone track *F* (Fig. 2), was made by a rectangular dolomite block, 11 by 6 by 5 inches, that weighs about 30 pounds. The writer's map, notes, and photos are supplemented by McAllister's prior photos of track in fresher condition for the following description. The track starts from a depression in edge of the chip litter, 5 feet north from a large rock (25-inch rock, Fig. 2). The sharp furrow extends N. 15° E. for 82 feet across clay, almost in a straight line, but with faint *S*- curvature, to the edge of chip litter at a blunt point of the shore. For a very few feet the rock grooved through the chips to clay, then continued for 37 feet across stones to clay again on north side of the point. In this interval the track is obscure, but the writer observed that it was marked by a line of small, up-ended stones, dislodged from normally smoother pavement. The stony point is 0.19 foot higher than the adjoining playa surface and 0.25 foot higher than the bottom of the furrow in clay, so the rock had to climb 0.25 foot in traversing the stones. On the north side of the point, the track is peculiar for 10–15 feet, partly contiguous indentations in an irregular line in the clay, partly poor groove; after this it gradually becomes a clear groove (Pl. 1, fig. 3) in 1–5 feet, toward N. 24° W., and then turns abruptly and extends N. 6° E. for 47 feet, where the groove abruptly terminates; a series of zig-zag indentations, like the previous ones but shallower, continue the track for 14 feet to the rock at the terminus. The total length is 194 feet.

Although most of track *F*, like all others on the playa, indicates sliding and not tumbling of the rock, its abnormal indentations are a puzzle. One geologist gave opinion that the rock, impelled directly by wind, slid on the clay up to the stones, then tumbled or rolled across the point to clay where skidding resumed. The writer has difficulty in conceiving that wind could tumble a rectangular block of this size any more than to upset it if standing on end. It would be contrary to laws of transport for wind to roll away this large rock and leave behind the hundreds of tiny stones that constitute the point traversed, even though most of the paved stones protect one another from wind. Also, to slide the rock over stones, unlubricated by wet clay and slightly upgrade, would require more force than wind could apply to it without ice as intermediary. The writer suggests the following. The rock was probably in an ice floe when beginning its track, but, after it reached the point, jerky movements from dislocating little stones caused either the rock to break loose from ice, or a small cake of ice holding it to crack away from the main floe, which then shoved, and possibly tumbled, the rock along fitfully. Once it had broken loose, its trail should be different from that made when it was firmly held by a considerable mass.

TABLE 5.—DETAILS OF LONGEST STONE TRACK, *D*

Stop	Leg	True azimuth (*in degrees*)	Distance (*in feet*)	Remarks
0	...	...	...	Conspicuous depression at south edge of playa
	1	10½	434	Well-indented track
0	...	...	...	Distinct but shallow depression
	2	183	205	East edge of track indented more lightly
0	...	...	...	Faint depression
	3	15	233	Well-indented track
0	...	...	...	Position of rounded dolomite boulder, 10 x 7 x 6 inches, weight about 20 pounds, at end of track
...	...	16	408	Net travel of boulder
...	...	...	872	Total travel of boulder

Machine-gun shells, cartridge clips, and barrel hoops have inscribed traces on the Racetrack, and the antiquity of some is limited to a decade. Wild jackasses have left their droppings, and these too have made tracks. A group of these is illustrated (Pl. 1, fig. 4) and strongly suggests identical signature for the four tracks. Still more significant is the obviously incised groove at near end of the right-hand track. If such light objects were blown separately by the wind and subject to tumbling, it seems unlikely that furrows would result; if the objects are frozen and attached to the mass of an ice floe the tracks are more logical.

Certain trails were not produced by integral scribers. Broad, scraped-over swaths ending in ridges of accumulated clay have not received as much attention as the rock trails. L. G. Kirk (1952a, p. 25; 1952b, p. 177) has illustrated a good example which is 84 feet long and widens from 2½ inches at start to 12 feet at a terminal ridge of mud flakes derived from the entire area of the swath. The writer is inclined to interpret these as scraped over by the leading tongue of an ice floe or small cake, which pressed more widely on the mud as it overreached the bounds of a shallow supporting pool. He has seen nothing to indicate that the mud here has been curled up and rolled along by the wind, snowballing, and accumulating moist clay, as has been suggested (R. E. Kirk, 1953, p. 323). One published photograph illustrates the same suggestive parallelism that many stone tracks do (L. G. Kirk, 1952b, Fig. 3, p. 177).

Evidence of Ice Action on Shore

Part of the playa shore revealed phenomena inexplicable except by ice action. On the east side of its southernmost arm, the steep hill slopes which undoubtedly supplied a high proportion of the wandering stones (Figs. 1, 2; Pl. 1, fig. 3) are separated from the flat, polygonally cracked mud by a littoral transition zone composed of (1) a thin pavement of small, angular rock fragments, an inch or so thick, over the peripheral clay; and (2) intermittent ridges or alignments of larger stones along its landward edge. The last are clearly small ice ramparts.

The pavement is so termed despite fine grain; its pieces are as closely fitted as shapes permit; flattish sides are generally up and flushed, and integrated into a smooth, level surface of mosaiclike texture. Its texture opens up and exposes clay toward its border with the playa, and on the landward side, parallel to the adjoining foot of some ice ramparts, the pavement is striped by windows eroded through it to clay. Innumerable faint stone tracks line pavement surface in various directions from onshore to alongshore, predominantly northeast. Except for a few that cut through to clay, one diagrammed (Fig. 2, *J*), most are inconspicuous or invisible in mid-day sun, but they multiply in cross lighting. Several overrun the mosaic and start up the hill slope. The oblique track at *H* (Fig. 2) climbs 1 foot in 18 feet to the stone (6 by 2 by 2 in.) at its end; the other three traces at *H* extend 12–15 feet with ascents of 1.2–1.4 feet to terminal stones of like size; one (*L*) rises 1.63 feet in 28 feet from pavement to 7-pound stone (7 by 5 by 3 in.) on hill slope and crosses an ice rampart enroute, as do others (*G*, *S*). Trails repeatedly score west-facing littoral for a mile to the northeast. Only ice seems capable of making them.

The ice ramparts are telltale despite unimpressive size. They stand a few inches above the rear of the chip pavement in even lines, and their aggregated coarse rock contrasts with the ground on either side. The pieces are angular and range up to 10 inches, but in almost every rampart they diminish in size northward. Many ramparts are buttressed at the south end by an unusually large rock or a hummock at a clump of sage.

Evidently ice floes have both a thrusting and a shearing action along this shore. Tracks at *J*, *H*, and *L* attest the former, as do three scalloped ramparts, the most southerly shown (Fig. 2). A more conspicuous effect is northward to northeastward scraping with enough eastward pressure to keep ice against littoral; and with longshore drifting of rock fragments. Stones were dragged northward occasionally scouring windows in pavement as they were strung along into rampart accumulations, or, where the shore recedes eastward, they embarked on a journey over the playa. Here undoubtedly is both the site and the means of launching most erratic limestone blocks and fragments adrift on the Racetrack. Some of the largest rocks are illustrated (Fig. 2).

The position of one rampart is like that of a spit (Fig. 2; Pl. 2, fig. 3). It starts from an 18-inch rock and contains stones up to 10 inches in its basal part, up to 8 inches at midway point, and up to 2 inches in distal portion. Once grown, the spit rampart probably deflected some ice floes, for opposite it track *E* (Fig. 2) curves gently in harmony, though tracks at *S* cross its end.

Negligible water transport or rounding of litter chips shows that shore is little affected by wave action. Many rocks strewn along ice ramparts are too large to have been moved

by the small waves of so shallow a lake. Altogether this particular reach of rocky shore exhibits more the action of lake ice than of lake water. Along the west side of the playa, less than a mile north of the south end, are three steel shafts driven in the littoral chip pavement. They protrude more than a foot above ground, 12–15 feet from one another, lean southward 30°–40°, and have been thrust out of vertically driven positions. Whatever their purpose, they have been disturbed by ice shearing southward along the west shore, in the same direction as most tracks on this part of the playa.

It is doubtful that ice is ever thick enough on the Racetrack to play any part by thermal expansion, and it is certain that the shearing action was by blown ice floes.

The evidences of northeastward force dominating in both ramparts and tracks near the south end, at least in mapped area (Fig. 2), and in spite of much greater playa fetch for nearly opposite movement, corroborate comments by Mr. Stough that such winds prevail here and analysis of winds by Shelton (1953, p. 438; Ubehebe Peak quadrangle sheet).

Discussion

Reports by Others

McAllister and Agnew (1948), during mapping in Ubehebe Peak quadrangle, briefly called attention to stone tracks on the Racetrack and "proposed that the scrapers were propelled over the muddy surface of the playa, initially from the rim, by strong gusts of wind blowing consecutively from different directions, such as the erratic whirlwinds that commonly produce dust-devils" (p. 1377).

L. G. Kirk produced quantitative information for various Race-track phenomena, including the task of weighing many larger rocks, and some excellent photographs (Kirk, 1952a; 1952b; Anonymous, 1952).

Some of his rocks (1952b, p. 176, 178, #7, #6 and #8) are shown on the writer's map (Fig. 2, 17″, 29″, and 22″ rocks, respectively, at ends of tracks *Q* and *R*). His trails #7 and #8 (1952b, p. 179) are the writer's *Q* and *R* (Fig. 2), and his data on course and length for them agree substantially with the writer's except that the two legs of Kirk's #8 are presented in transposed order; both *Q* and *R* have been illustrated previously (*Q*, L. G. Kirk, 1952a, p. 24, upper left, and R. E. Kirk, 1953, p. 321, lower right; *R*, L. G. Kirk, 1952b, p. 180, Fig. 4, R. E. Kirk, 1953, p. 320). Track *Q* (Kirk's #7) in some illustrations (Fig. 2; Anonymous, 1952, p. 53) has a northward extension of about 34 feet added to the original length of about 210 feet. The extension is slightly offset and fresher than the longer part; and different grooving shows that the stone was either reoriented or resting on a different face. Reliable report indicates it was made by some persons dragging it behind a truck, and this is the only artificial stone track here known to the writer. Kirk's trail #4 is located by writer at *A* (Fig. 1) and illustrated (Pl. 1, figs. 1, 2; also R. E. Kirk, 1953, p. 322, upper left). Kirk reluctantly accepts the wind theory as the "best theory to date" (1952a, p. 25).

Ruth E. Kirk points out that the Racetrack trails are much longer than the ice-shove furrows generally observed about lake shores. This is true, but the writer does not accept this as support for the conclusion that ice did not make them. The difference is certainly no greater than the contrast between the level floor of Racetrack playa and the sloping beds of most lakes where ice-shove marks are recognized. The length of playa inscriptions is due to the distance a floe may drag on bottom in uniform depth of water.

Shelton (1953, p. 438) demonstrated on the Racetrack that it is possible for propeller wash of an airplane to move small stones on wet clay:

> "two natural stones (found on the playa), including the 19 ounce limestone ... moved several times over distances up to about 2 feet under measured wind velocities of 40–45.5 m.p.h. ... Movement was slow and halting, and the mud surface was actively rippled by the air blast. ... under the conditions of the experiments any tracks formed might have been blurred or destroyed by further action of the wind. ... (with) laboratory-cut cubes of limestone in graduated sizes, ½ to 3 inches on an edge, ... none of the prepared cubes slid though the smaller ones showed some tendency to roll over."

No tracks are reported where artificially made wind did move small stones over water-saturated clay, and cubic stones tend to tumble.

Clements (1952) found stone trails at Little

Bonnie Claire Playa (elev. approx. 4000 feet), Nevada. On his second visit within 6 weeks he found new trails added by a 24-hour storm which had brought snow on the surrounding mountains, with: "snow line only slightly above the level of the playa" (p. 186). The playa was covered with water not over half an inch deep; the numerous trails ranged up to 138 feet long and were made by stones with dimensions up to 6 by 4 by 2 inches. His conclusions, that these trails: "have been formed by rocks pushed over the wet and slippery surface by the wind, . . . movement by ice, seems to be ruled out" (p. 186) do not convince the writer, even though small tracks were made near large stones which did not move at the same time. Indeed his statement of weather and the curved and seemingly parallel paired trails (1952, Pl. 1-B) suggest ice was the agency.

A significant point concerning the scarcity of playas with stone trails is brought out by a question, quoted but unanswered in Clements' paper (1952, p. 182).

> "One of the questions that has consistently been asked when the wind has been suggested as the motivating force of the moving rocks is: 'Why are such furrows not reported from other playas?' . . . casual personal observation of a number of other playas had brought to light no similar trails."

A logical answer might be that most other playas (of Mohave-Colorado desert) are lower, and temperatures to develop sufficient ice would depend considerably on elevation. Also, loose rocks must adjoin the littoral to be within access of floe ice. A few tracks are reported by Brigham A. Arnold (Personal communication) on two playas (Laguna Seca) in the Chapala basin, elevation about 2000 feet (?), north latitude about 29½°, Baja California. Recently John Shelton (Personal communication) has discovered a single track made by a 1½-pound piece of basalt moving 35 feet northward, near the west side of Nelson Dry Lake, at an elevation of about 1900 feet, 8 miles southwest of Boulder City, Nevada.

Ice Floes and Weather Considerations

On December 3, 1952, the central transcontinental telephone line east of Reno, Nevada, was suddenly disconnected for unknown causes (Roscoe M. Smith and L. L. Parish, Personal communication), and a repair crew was sent out to find the difficulty. After following the line eastward from Reno the crew came to Toulon Lake, a part of Carson Sink covered with a year-round body of shallow water.

Toulon Lake (elev. 3895 feet) is 6 miles long and ¼–¾ miles wide, parallel to the west side of Humboldt Lake and less than 1 mile from it; it does not show on the Lovelock or Carson Sink quadrangle topographic sheet (U. S. Geol. Survey, 1910) but has evidently been filled recently by irrigation and drainage rearrangements of the Humboldt River in the agricultural area around Lovelock.

At the lake, partly ice-covered, the repair crew discovered that 3000 feet of line was missing; this included 20 poles loaded with 4 arms and 40 wires on 150-foot spans. Some poles had been set in caissons of corrugated metal 10 feet in diameter and 4 feet high, filled with boulders to protect the poles against waves on the lake. Ice had formed for 3000 feet from the west shore and 2 or 3 miles along it, with open water to the east; the ice was 4 inches or more thick and floated on about 2 feet of water. Emergency repairs were made by dragging an insulated cable across the ice with a propeller-driven ice boat. The uprooted poles and remnants of wrecked caissons were found 300 feet south of the original position of the line. After two days the wind reversed direction, and the ice started to move north, tearing out the emergency repairs; it continued to move north under the influence of a 30-mile-an-hour wind a total distance of 700 feet and left the wrecked poles and caissons 400 feet north of the original line. Ice was pushed up several feet on the east shore[1].

The preceding demonstrates force sufficient

[1] While this paper was in press the writer visited Toulon Lake to look for tracks made by the dragged poles, etc. Only obscure or indefinite trails, from stones, timbers, or boards, were to be seen on the lake bed and its east shore slope. The silt-sand pulverizes easily and would not retain tracks as well as Racetrack Playa. Vernon Cannon, employee at Cooney's Mill near by, informed the writer however that he had once seen two tracks there about 40 feet long made by stones (10–12 pounds) in the southwest ends. According to him ditch changes filled the lake first in 1942 and emptied it in 1954. Mr. Cannon also mentioned his observation of a track made by a very large (4 feet) rock on Black Rock Desert (northwestern Nevada).

to move large rocks over shallow lake beds, and the writer maintains that ice furnishes the only feasible force. Another item is also significant. Mr. Parish (Personal communication) suspects that wind tended to lift the poles by driving water under the ice floe which clung to poles and swamp braces. He cites similar action on Grass Lake (Siskiyou County, California) where rising water, buoying ice, jacked up poles about 1 foot, and the poles later sank back after ice was chopped away. Toulon Lake was not raised by floods at the time of the break, for Humboldt River flow was at a minimum. Ice floes on water raised either by wind or inflow could easily ferry off rocks without making tracks.

Ice-floe activity on Racetrack playa requires three weather elements—precipitation, freezing, and wind—in proper and close succession. From various reports, water occurred on the Racetrack three times during the winter of 1951–1952 and twice the following summer, probably more often than usual. Mr. Crook, here many years, reports that a little water entering at the north during calm weather gradually works to the south end. The deepest water observed by him was 10–12 inches; the shore lines observed by the writer indicate no greater depth. Significant for part of the writer's interpretation is Crook's report that wind frequently drove all water to one end of the playa and kept it there for the duration.

Ice has been seen on Racetrack playa by Joseph Crook (1906–1907, 3/8 in. thick) by McAllister (1947–1948, repeatedly, presumably enduring several weeks), and by Stanley Jones (about 3 inches thick, late January 1949). Jones drove a pick-up truck over it without breaking ice which rested directly on clay. Such thick ice probably occurs seldom here, for this was after a freak storm which not only closed all passes but brought 4 inches of snow to the floor of Death Valley. Blowing over water, this ice could have moved the largest rocks on the playa. Thinner ice certainly forms more frequently; George Stough reported almost yearly trouble from frozen water lines at the Lippincott Lead Mine.

On the frequency and times of track formation, the writer can only estimate for the Racetrack; that small stones are probably moved several times, large rocks perhaps once per decade; that the deep furrows *Q*, *E* (Fig. 2), and *A* (Fig. 1; Pl. 1, figs. 1, 2) were probably formed in the period 1947–1949, *D* and *R* (Fig. 2) at about the same time or earlier, and the signature groups *A*, *B*, *C*, etc., (Fig. 2;

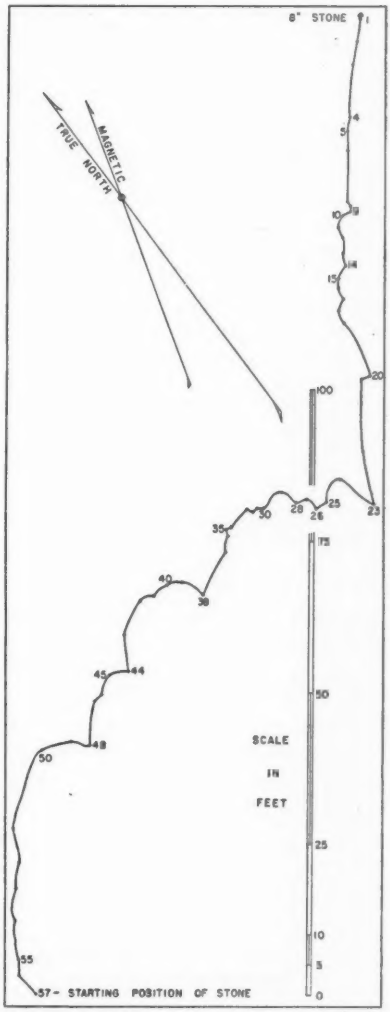


FIGURE 5.—MAP OF STONE TRACK IN NORTH PART OF PLAYA

The track was formed probably in storm of late January 1954. Its irregularity contrasts with others mapped. Within 100 feet northwest, and 300 feet southeast of it, three additional tracks had similar signatures and direction.

Pl. 3) and *K* (Fig. 5) probably later, 1950–1952. Obliteration of tracks is evidently irregular, for the writer has observed some dimmed but others unchanged between visits. Kirk (1952b, p. 178) reports similarly.

On lower playas, though freezing would not be expected so frequently, temperatures appropriate for it occur. At Las Vegas (with elev. 2033 feet, almost the same as Nelson dry lake) minima of 12°F. (Dec.), 8° (Jan.), and 10° (Feb.) have been recorded (U. S. Weather Bureau, 1932b); at Blythe (elev. 268) of 11° (Dec.), 5° (Jan.), and 19° (Feb.), and in the bottom of Death Valley (elev. −180) of 19° (Dec.), 15° (Jan.), and 27° (Feb.) (U. S. Weather Bureau, 1932a).

During the winter of 1915 Mr. Crook showed stone tracks to his wife who could not believe the larger rocks would shift naturally. Accordingly she marked the position of a big one which eventually moved away.

New tracks were formed by small stones in the winter of 1953–1954. Though skeptical of ice-floe origin, George Stough (at Lippincott Lead Mine, 4 miles south of Racetrack and slightly lower elevation) had agreed to look for fresh tracks. He wrote in February 1954:

"We had a big storm just before I left (about Jan. 25) ... three hours drizzling rain, then about five inches of snow ... more in the hills. The Racetrack was completely covered with water when I left and snow still melting. Freezing every night. Will check and let you know if there was any action. Would like to have stayed and seen any action as it took place".

He wrote again 3 days later after searching the playa:

"there are several new tracks across the lake that were not there before. Looks like you win. ..."

The author re-examined Racetrack Playa in mid-April 1954 and found the following fresh tracks which probably were formed in late January 1954; (1) About 2200 feet east–northeast of the Grandstand, a series of six widely spaced, similar, zig-zag tracks all running northeast; the longest showed a net travel of 170 feet, N. 55°E. of the 8 by 4 by 3-inch dolomite rock at its northeast end, but it measured 223 feet along its irregular course (Fig. 5), (2) near by, three similar tracks, probably of same signature; (3) a few tracks near the south end, with northeast movement; one was only a yard from the stony pavement of the shore, and its trend was parallel to shore. This 2-inch-wide track contained many fine, parallel striae made by surface irregularities of the 2-inch stone which lay at its northeast end in the time-softened foot imprint of a wild jackass. The stone had not crossed this depression in the clay but had come to a stop on the downslope where it entered it. Another fresh track (*Y*, Fig. 2), much longer and narrower and made by a smaller stone, passed within 20 inches of the rounded 10-inch stone, unmoved, at the end of the older track *D*. It started northwest, circled sharply left, and headed north across its own track and then by the larger rock. Neither of these fresh tracks was present in November 1953. (4) Along the west side of the playa about 100 fresh tracks were seen. All showed southward movement, like most older ones here. They ranged from a few inches to more than 20 feet long, but within a few acres both the lengths and patterns of numerous tracks were much alike. Small stones, and in some cases twigs of sagebrush, were at the southern ends. In one case a small, fresh track ended at the older, deeper, and broader furrow made some years before by a large rock. The small stone had entered the depression and there stopped, and its track did not continue across the older one, though the clay surface immediately ahead was smoother than where it had just traveled. This is similar to performance of the 2-inch stone in (3). (5) Southeast of the Grandstand, over the central northern part of the playa, many tracks showed where small stones, or burro droppings, had traveled 27–75 feet south-southeast.

Scriber shifting and the premature end of tracks *A*-5 and others indicate that small stones were sometimes insecurely held by moving ice. Small stones would not project much above the mud surface, and a floe in shallow water, riding and rubbing over them, might attach and move and later release them. The sudden stops with increased depth of water, even on a down gradient, suggest that the stone's weight was sufficient force to detach it; or, for other reason, it was disengaged when no longer squeezed between ice and clay. Water-level changes due to wind could aid in attaching or releasing scribers and might determine the start of small

tracks or the end of stoneless ones. Stones in shallow water might be pushed by the forward edge of a floe without being fixed to it (parts of *F*, Fig. 2). Thin ice would tear away from large rocks while removing smaller ones.

Another possible case is a thin ice floe, barely floating because of shallow water, pushing its forward edge over moist clay that might be (1) saturated without appreciable depth of water over it, (2) lightly frozen on the surface, or (3) unsaturated and firm. The mud trails described seem a likely product of conditions (1) or (2). Condition (2) might supply the mud flakes described by Shelton (1953, p. 438). Condition (3) is the only explanation of several patches of clay near the south end of the Racetrack examined in April 1952. The Racetrack had recent water (puddle, Pl. 1, fig. 2). The clay for several yards in all directions was covered with fine parallel striae and a slight burnish. There were no stone marks. Ice must have scraped the clay which could not have been saturated; it might not have been even moist.

Competent geologists, familiar with the writer's data on stone tracks, maintain contrary opinions about the origin of mud trails and believe they are formed directly by wind. It has been considered impossible for ice to form one particular mud trail. The writer tentatively offers the following explanation. He believes it feasible for an ice floe to push several thin ice cakes before it, like an Ohio River work boat and scows in front, or for the leading tongue of a floe to break into separate cakes. In either case, after scraping mud from trail into a ridge at the leading edge, separate cakes could trace spur trails, branched away from the trunk one, after wind changed. Kirk (1952a, p. 25) described three offshoot trails which turned and paralleled the main swath, and finally merged.

Stones overlying discrete mud flakes at ends of stone tracks (Shelton, 1953, p. 438) and stones on terminal clay mounds at ends of mud trails (Kirk, 1952a, p. 25) are also peculiar. For the latter case, a stone might be snowplowed up onto the edge of an ice cake and left partly on terminal mound when ice melted. This is more likely than for a wind-gathered mud mound to remain while a 6-inch rock was independently blown onto it. The mud flakes are less easily explained; the writer suggests they may have been frozen, and Shelton thinks they were nearly dry or held by algal filaments. The writer believes that: (1) mud trails show scraping as by a broad, solid edge, and some grouped ones exhibit parallelism and similar bends, (2) wet mud could not be scraped up and accumulated into mounds by wind, and (3) dried mud, grains, flakes, or curls could not be so selected by wind as to produce the sharp-bordered, cleanly swept trails; that it would be blown into even, terminal ridges is also doubted.

One final, speculative comment arises from careful comparison of photos of track *Q* (Fig. 2, natural part only) and that at *A* (Fig. 1). Photo illustrations for both have been fully referenced. Though over half a mile away and several times longer, *Q* appears to match closely with the track at *A* which the writer suggests may be its telescoped counterpart. Three tracks are needed to test an ice floe's motion. If a third similar one is found elsewhere, a test would be possible and might prove signature identity. If so, this floe was much longer than any previously postulated by the writer, and, inasmuch as between the rocks of these trails are larger ones not so moved, a necessary conclusion would be that the intermediate rocks cut long, interior slots or hollows in moving ice.

Conclusions

(1) Certain groups of stone trails on Racetrack Playa have identical signatures over areas nearly 500 feet across. They must have been formed by blown ice floes which held numerous small stones or scribers in generally fixed positions, and which rotated during travel.

(2) In several cases a single, moving ice floe released a small stone and picked it up again after a few feet of travel.

(3) Many tracks favor origin by ice floes blown across the playa by wind rather than by wind blowing the lone objects.

(4) Mathematical treatment applied to stone trails within a signature group demonstrated that the scribers moved in accordance with trigonometric relationships between separated loci in a rotating, planar body.

(5) There are ice ramparts and other signs of ice-floe action on Racetrack Playa, aside from the stone trails.

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DIRECTIONAL-CURRENT STRUCTURES FROM THE PREALPINE FLYSCH, SWITZERLAND

By John C. Crowell

Abstract

Gurnigel sandstone beds in the Paleocene-Eocene Flysch of the External Prealps, Switzerland, contain primary structures which display common orientation through many feet of strata. The linearity of some of these directional-current structures is easily measured in the field. The bottom of beds display groove-cast and load-cast primary-current lineations; within a layer are found clast lineations subdivided into grain and charcoal-fragment lineations, and parting lineations consisting of streaks on bedding-planes. Flute casts, torose load casts, current stratification, convolute bedding, and ripple marks also occur. These, and other factors, suggest that the sediments were laid down in a relatively sheltered basin by turbidity currents.

In three sections of the Gurnigel Flysch 401 field measurements of the orientation of current structures show that during the Paleocene and Eocene a source for sediment lay not far to the northwest. This information, with published data from clast types and the distribution and facies of other Paleocene and Eocene units, implies that a land-mass or island probably lay within the northwestern part of the Ultrahelvetic sedimentation region within which the Gurnigel Flysch accumulated, and near the border of the Helvetic region. The Oligocene and Pliocene culminations of the Alpine orogeny carried the Gurnigel Flysch far to the northwest within Ultrahelvetic nappes.

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Introduction

Sandstone layers in the Flysch of the External Prealps, Switzerland, display numerous small-scale sedimentary structures, such as current bedding, groove casts, and load casts, which were formed when the strata were deposited. Inasmuch as these structures show a remarkable degree of common orientation through many feet of strata, they may be useful in determining the nature and direction of the currents responsible for them. They are therefore described and classified here, particularly from the viewpoint of the field geologist interested in obtaining rapidly some information on the slope of the sea floor when the sediments were laid down, and the direction to the source of sediment supply. Data gathered within the External Prealps and other facts have forwarded understanding of the paleogeography during deposition; an effort has been made to fit these data into present concepts of the geologic history and tectonic emplacement of the Prealps.

Previous workers have also employed sedimentary structures to determine current direction and sense in the area of deposition. In 1843, Hall (p. 234) used groove and flute casts as indicators of current directions in the Devonian of New York, which has been studied more recently from this viewpoint by Clarke (1918) and Rich (1951, p. 12). Cross-stratification was used as an indicator of current sense in 1858 by Sorby (1908, p. 197), and has since been studied by Knight (1929), Brinkman (1933), Shotten (1937), Jungst (1938), Reiche (1938), Bausch van Bertsbergh (1940), McKee (1940), Wilson *et al.* (1953), Potter and Olson (1954), and others, as have ripple marks by Hyde (1911), Bucher (1919), Kay (1937; 1945), Bausch van Bertsbergh (1940), and King (1948). Oriented fossils, such as graptolites, cephalopods, and fusulinids, were used by Ruedemann (1897), Kay (1945), and King (1948). Many of these papers contain maps with current arrows. In Germany, Richter (1935), Häntzschel (1935), Kraus (1935), Kieslinger (1937), Rücklin (1938), and Bausch van Bertsbergh (1940) investigated many of the kinds of structures described in this paper, and other workers have also contributed to an understanding of these structures (Fuchs, 1877; 1895; Fugger, 1899; Shrock, 1948).

Kuenen (1952; 1953a; 1953b), Kuenen and Carozzi (1953), and Kopstein (1954) have pointed out the ease with which information on current direction and sense can be obtained by noting the orientation of small-scale sedimentary structures. The work of Kuenen has given new insight into the way that certain sequences were laid down by turbidity currents, the significance of which had been largely overlooked by geologists and oceanographers. Our understanding has come about through studies of sedimentation processes in the laboratory (Kuenen and Migliorini, 1950), through the recognition of sand layers in relatively deep basins of present-day oceans (Shepard, 1951; Erickson *et al.*, 1951; 1952; Ewing, 1953, etc.), and through the study of sedimentary rocks in the field (Natland and Kuenen, 1951; Kuenen, 1953a, 1953b; Kopstein, 1954).

The writer is very grateful for the Guggenheim Fellowship and the Fulbright Award which have made his Alpine studies possible while on sabbatical leave from the University of California, Los Angeles. He also thanks Professor Bruno Sander, University of Innsbruck, Austria, for the use of space and equipment in his Institute where the laboratory and office work were undertaken. Appreciation is extended to Professor Jean Tercier, University of Fribourg, Switzerland, for critical reading of the manuscript and for illuminating discussions on the geology of the Prealps. The writer is also indebted to Edward L. Winterer and Ph. H. Kuenen for comments on the manuscript and to Betty Bruner Crowell for the drafting.

Description of Sections Studied

The three sections of the Paleocene-Eocene Flysch studied for this paper are exposed in the northeastern arc of the external Prealps between Lake Geneva and the Aare River in western Switzerland (Fig. 1). They lie near the northwestern limit of the Ultrahelvetic Nappes, and just above the complex of southeast-dipping thrusts which marks the base of the giant Prealpine klippe. Large-scale sketch maps of the three sections are shown in Figures 2 and 3 and the local geology on Figure 4.

The Flysch rocks of the External Prealps, which reach a thickness of about 3000 m, consist of two main stratigraphic parts with an

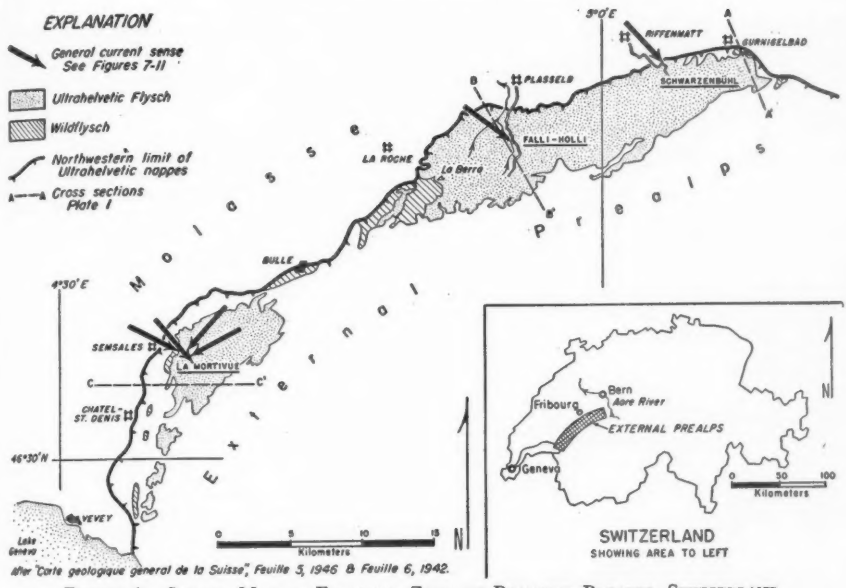

FIGURE 1.—SKETCH MAP OF EXTERNAL ZONE OF ROMANDE PREALPS, SWITZERLAND
Location of the three sections studied, and the general current sense as inferred from primary sedimentary structures, are shown

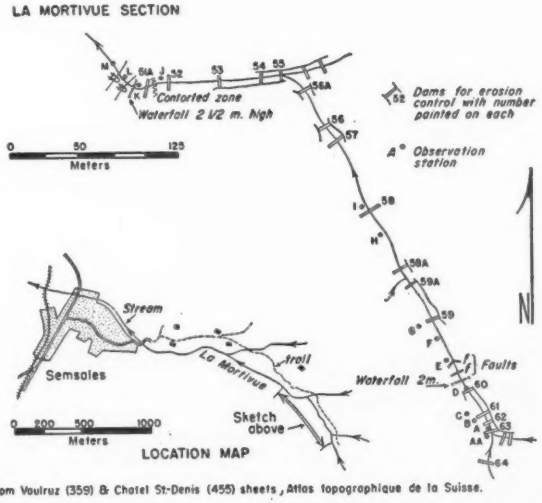

FIGURE 2.—SKETCH MAPS SHOWING LOCATIONS OF OBSERVATION STATIONS IN THE LA MORTIVUE SECTION

interfingering gradation between: the Wildflysch below and the Gurnigel sandstone above (Tercier, 1928, p. 3, 65). The Wildflysch, of variable character and thickness, is composed of marly shale with interbedded sandstone, breccia, conglomerate, and limestone. but is best known for its exotic blocks and thrust slices in the shale. The Gurnigel sandstone,

with which we are concerned here, is made up primarily of interbedded sandstone and shale with some limestone and minor breccia and conglomerate. On the basis of *Nummulites* from the limestone beds it is assigned to the Aurversian and Lutetian stages of the Middle Eocene and extends downward through the Ypresian (Lower Eocene) and into the Paleocene (Tercier, 1928, p. 62; Cadisch, 1953, p. 176; Tercier, 1954, personal communication). The Flysch sandstone (Flysch gréseux) which overlies the Wildflysch in the vicinity of the La Mortivue section is on the whole equivalent to the Gurnigel sandstone but may range up into the base of the Priabonian or Upper Eocene (Gagnebin, 1924, p. 32).

Exposures of the Gurnigel sandstone consist of gray-green sandstone layers, weathering gray brown, interbedded with dark shale (Pl. 1, fig. 1). The flaggy sandstones range in thickness from about 1 cm to more than 2 m, but most are from 10 to 20 cm thick. The proportion of sandstone to shale in an exposure ranges widely, from about 30 per cent sandstone to about 80 per cent, with the thicker and coarser units where sandstone predominates. These layers are much more resistant to weathering and erosion than the intercalated shale, which

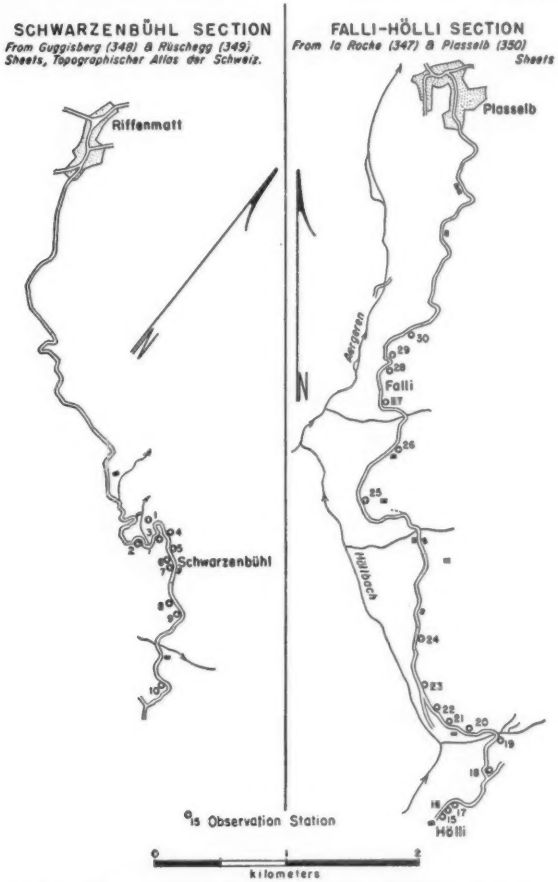

FIGURE 3.—SKETCH MAPS SHOWING LOCATIONS OF OBSERVATION STATIONS IN THE SCHWARZENBUEHL AND FALLI-HOELLI SECTIONS

FIGURE 4.—GEOLOGIC CROSS SECTIONS THROUGH EXTERNAL PREALPS

breaks into minute shards, so that the shale can be studied only where outcrops are particularly good. The formation is poorly exposed except in a few road cuts, canyons, and quarries, and underlies rolling grassy and wooded hills. At several places the sandstone is quarried for paving blocks and building stone and in fact the writer first noticed the current structures described here in curbstones of near-by cities.

Many of the sandstones are more properly subgraywackes and arkoses since feldspar is nearly everywhere present and commonly predominates. Impure calcarenites, which consist of sand-sized angular shards of limestone and other grains mixed with considerable organic debris, are also common. Glauconite, mica, altered amphiboles, and other dark minerals are conspicuous among the grains of quartz, feldspar, and fragmental limestone. Grain sizes in all types range from very fine to coarse, with fine the most common. Beds are poorly sorted, and a large proportion of silt and clay is found in the matrix along with ubiquitous lime cement, which is commonly stained with iron. The rocks are well cemented and indurated so that intricate cast structures have been preserved on the bottom of beds. Joints, many filled with calcite, cause the beds to break upon weathering into angular blocks. In places the layers are cut by numerous closely spaced joints so that a lineation is formed on bedding planes which needs to be discriminated carefully from primary-current lineations. The petrography of the different rock types within the Gurnigel sandstone has been investigated by Gerber (1925, p. 21) and Tercier (1928, p. 48).

Tracks of organisms and burrow-fillings are quite conspicuous on the bottoms of many sandstone layers. Three types have been recognized: (1) Shallow grooves up to 1 cm in width may represent tracks, perhaps of a gastropod. Groove-cast lineations pass through these tracks and are therefore later. (2) Depressions, perhaps siphon holes and certainly organic in origin, are filled in places with the overlying sandstone which came in after they were formed. Some flute casts commence at these depressions so as to suggest that they were influential in localizing the flutings, and a few deep load casts have settled into them. (3) Most of the conspicuous organic trails must be burrow-fillings which were formed after the overlying sandstone was laid down, as the vermiform tubes are filled with reworked sand from the overlying bed. At places these are so numerous that most original sedimentary structures have been destroyed.

The dark pelitic rocks are both laminated and massive, and are found in beds up to a few centimeters thick which in many places possess no internal structure discernible in the field. These massive rocks (mudstone, claystone, and marl), without fissility or lamination, and composed of clay, silt, or very fine-grained limestone detritus, contrast with laminated and fissile shale. Shale beds commonly show varvelike graded layers with two parts: a lower coarse unit with siltstone (or very fine sandstone) at the base which grades upward in less than 1 cm to a claystone or marl unit several centimeters thick. Thin laminations within the lower part are made conspicuous by slight changes in grain size and color. Here large flakes of biotite and flecks of charcoal or carbonized wood, many with a preferred orientation, may lie within the laminations. The thin layers of very fine sandstone commonly exhibit faint current stratification and vary in thickness across an exposure.

The Schwarzenbühl and Falli-Hölli sections on the northeast are relatively undeformed, and consist of gently folded or uniformly dipping strata. Only minor faults and a few small chevron folds with approximately horizontal axes were noted. The La Mortivue section on the southwest, however, is largely overturned, and several faults are marked by zones of sheared rock several meters wide. In addition, there are a few small folds with axes which plunge gently. Bedding planes at La Mortivue have therefore been rotated through the vertical from their original nearly horizontal position, and have almost certainly not been rotated around axes now parallel to the present strike. In the absence of a detailed structural map of the external Prealps, which will probably never be made in view of poor exposures, it is impossible to determine just how the beds have been rotated. Since no other technique seems available, directional observations measured in the field have been rotated around the present strike in order to bring them approximately into the position they must have had when formed. This technique is no doubt reasonable for the north-

eastern two sections, but an error of unknown magnitude (probably small) is contained within the rotated observations of the La Mortivue section.

Directional-Current Structures

Lineation

Small-scale sedimentary structures of the sandstone layers are among the most noticeable features of the Flysch rocks of the External Prealps. The undersides of beds may reveal intricate details where soft shale has weathered away. These sculpturings are commonly the casts of structures formed by cutting or sinking into mud below and now filled by cemented sandstone. Details are rarely found on bed-tops because of gradual grading from sandstone to shale. Many beds also display internal structures such as current and convolute bedding.

Many features described here are varieties of "flow marks" (Richter, 1935; Rücklin, 1938; Rich, 1950; Kuenen, 1953a) and the primary-current lineation is similar to that described by Cloos (1938) and Stokes (1947). Such lineations were produced experimentally by Sorby in 1849 (1908, p. 188), which he called "graining in the line of the current." Ruedemann (1897, p. 380) observed and measured the orientation of lineations in the Ordovician rocks of New York, although he used the term "driftline."

Because most of the current structures are arranged in lines or have elongated shapes, they display a marked linearity on bedding planes, which makes their measurement possible. Where the term *lineation* appears in the name, the linear nature on the stratification plane is a defining or striking characteristic. Other forms, such as flute cast, torose load cast, and ripple mark, may as individuals be elongate, but their linearity is much more impressive where many forms occur together. Not all other structures, such as current stratification and convolute bedding, display a linearity and some may require a three-dimensional view for interpretation as they are found within a bed rather than on a bedding plane. Most features with linearity indicate the strike of the current line but not the sense of current flow along this line; that is, a north-south lineation gives no information on whether the current came from the north or from the south. On the other hand, current stratification provides an approximate sense readily but a precise bearing or azimuth only after painstaking work. Where lineations and current stratification exist in the same bed, accurate bearings on the current line can be obtained quickly from the lineation, and the sense of flow along this line by a brief look at the cur-

Table 1.—Current Structures from the Prealpine Flysch

I. *On bottom of bed:*
- A. *Substratal lineation*
  - 1. *Groove-cast lineation*
    Grooves originally gouged or cut into top of mud layer and preserved as casts on bottom of overlying sandstone bed; straight continuous ridges, usually smooth; when formed, underlying mud was relatively hard. L ‖ St
  - 2. *Load-cast lineation*
    Fine-textured rills and ridges pressed into soft mud and preserved as casts on bottom of overyling sandstone bed; curved and braided discontinuous ridge, usually ropey; when formed, underlying mud was relatively soft. L ‖ St
- B. *Flute cast*
  Cast of fluting cut into mud layer. L ‖ St
- C. *Torose load casts*
  Ropey subconical load casts pressed into soft mud with long axes subparallel to stream lines. L ‖ St

II. *Inside bed:*
- A. *Internal lineation*
  Lineations on lamination planes within sandstone layers.
  - 1. *Clast lineation*
    - a. *Grain lineation*
      Elongate grains lined up parallel to flow. L ‖ St
    - b. *Charcoal-fragment lineation*
      Aligned elongate flecks of charcoal or carbonized wood. L ‖ St
  - 2. *Parting lineation*
    Faint irregularities or streaks on lamination planes. L ‖ St
- B. *Current stratification*
- C. *Convolute bedding*
  Axes of minor folds within convoluted beds roughly normal to flow or bottom slope. L ⊥ St

III. *On top of bed:*
- A. *Ripple marks* L ⊥ St

rent stratification. Flute casts in many cases give the sense as well as the direction of current flow.

For some structures the alignment is parallel to the current direction and for others normal to it. In Table 1, therefore, L (for lineation or linearity) and St (for current; stream or *strömung* in German) have been used in order to provide a short-hand designation of these relations. Some rocks may show two or more lineations, as for example where primary-current lineation is associated with a secondary lineation owing to the intersection of very closely spaced joints with the bedding plane in the La Mortivue section. Here lineations may be designated as L_1, L_2, L_3, etc.

Current Structures on Bottom of Beds

Substratal lineation.—The substratal lineation includes groove-cast lineation and load-cast lineation.

GROOVE-CAST LINEATION: The bottoms of many sandstone layers show numerous subparallel straight ridges with a relief of from less than 1 mm to more than 1 cm (Pl. 1, fig. 2; Pl. 2, fig. 1). At places the whole surface is striated although elsewhere a space up to several centimeters may separate striae. Most lines, which may extend for more than a meter, show a definite preferred orientation, but some depart from the average direction by as much as 15° to either side. Large groove-casts, commonly remarkably straight, show finer groove casts superimposed upon them. In cross section many are asymmetric, and on any one specimen not always on the same side. Under the 50-power microscope no dimensional orientation of the grains associated with the lineation was conspicuous although one might be revealed by a painstaking statistical study based on grain measurements.

The grooves were obviously cut into a clay or mud layer by a current laden with sand grains, pebbles, or bits of other material. However, in only one place was a lamination in the underlying shale distinctly cut by a groove. After the cutting, sand or silt filled in the sculptured lines so that they are now preserved as casts on the bottom of the overlying sandstone layers. The mud or clay was indurated enough to allow cutting rather than the formation of load casts, but it is not known whether the same current that did the cutting also laid down the sand. Groove lineations have been produced experimentally by allowing water with some sediment to flow over mud (Rücklen, 1938, fig. 10).

The term groove cast comes from Shrock who has included excellent illustrations from the Devonian Portage sandstone of New York and the Mississippian Horton sandstone from Cape Breton Island, Nova Scotia (1948, p. 162-165, figs. 121, 122).

LOAD-CAST LINEATION: The bases of many sandstone beds display a pattern of fine-textured rills and ridges (Pl. 1, fig. 4). The lineation appears braided and individual ridges are ropey, kneaded, or bumpy. Ridges extend for only a few centimeters at most, curve and bend, diverge and converge, and exhibit marked changes in longitudinal relief. Segments of ridges may depart as much as 30° from the average trend, but the average direction is distinct and easy to measure in the field. In cross section the ridges are usually steep-sided with rounded tops, and the troughs between crests have nearly the same shape in reverse.

PLATE 1.—GURNIGEL SANDSTONE AND CURRENT LINEATIONS

FIGURE 1.—GURNIGEL SANDSTONE, LA MORTIVUE SECTION, STATION AA

Width of exposure is about 7 m.

FIGURE 2.—GROOVE-CAST LINEATION

Note subparallel orientation of casts, large and small, on base of this overturned bed. Arrow shows sense of current. Scale at left is 20 cm long. La Mortivue section, station AA-1.

FIGURE 3.—PARTING LINEATION, LA MORTIVUE SECTION, STATION D

Note that the shadows reveal three bedding planes, each of which displays good parting lineation, parallel to the arrows (drawn on specimen in field.) Scale in centimeters.

FIGURE 4.—FINE-GRAINED LOAD-CAST LINEATION

On bottom of bed from La Mortivue section, station M-2. Arrow shows current sense. *Note* also burrow-fillings (bf) and tracks (t). Scale in centimeters.

FIGURE 2

FIGURE 4

FIGURE 1

FIGURE 3

GURNIGEL SANDSTONE AND CURRENT LINEATIONS

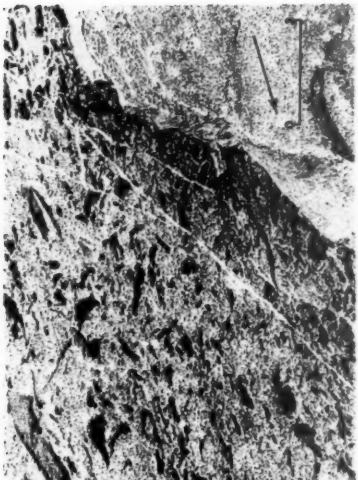

FIGURE 2

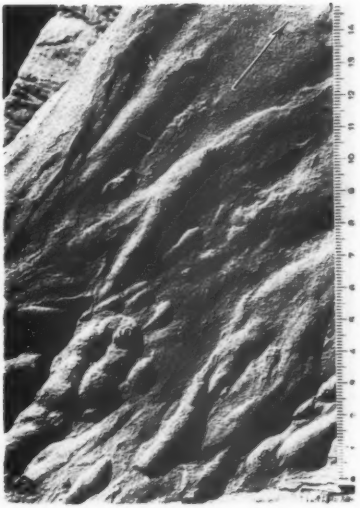

FIGURE 4

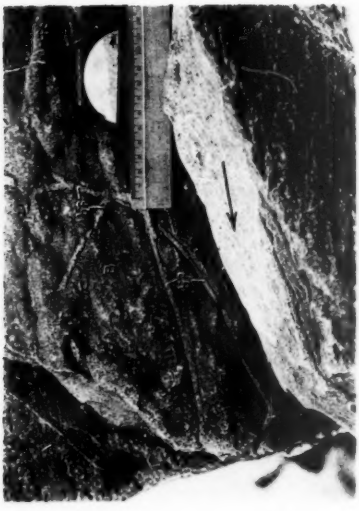

FIGURE 1

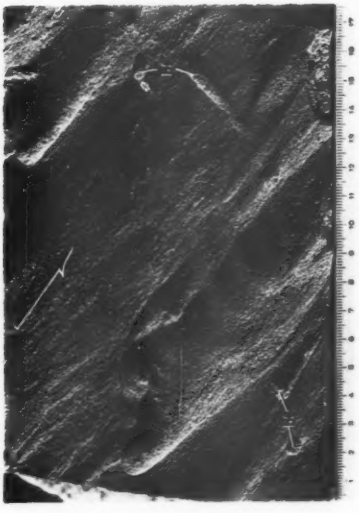

FIGURE 3

CURRENT STRUCTURES

At places larger rills and ridges may be finely etched with smaller sets, several to the millimeter. Some specimens exhibit low humpy load casts, a centimeter or so in breadth and up to a half centimeter in relief, with the rill and ridge pattern superimposed thereon. Many of these larger nonlinear load casts contain coarser sand grains, which shows that they were formed before or during the sand deposition, and a few of the linear load casts also contain coarser grains.

Load-cast lineation probably originated by the sinking-in of sediment emplaced by a turbidity flow in the manner described by Kuenen and resulting in the structure named by him load cast (1953a, p. 1058). Well-developed load casts are present throughout the three sections, and some large ones have clearly been pressed into underlying mud as shown by the deformation of the underlying laminations. Here the laminations were bent by sinking lobes of sand rather than truncated or cut by erosion along rills. Perhaps the underlying mud was relatively soft and pliable in comparison with indurated mud required for groove formation. Contrast in current velocity in addition to mud induration may have been an important factor: a sluggish and dense turbid current from which sediment was settling might form load-cast lineation whereas a fast-moving and unsaturated current would still be able to scour vigorously.

Flute casts.—Sharp subconical welts are commonly associated with groove-cast lineation (Pl. 2, fig. 3). One end of the ridge is rounded or bulbous, and the other flares out to merge gradually with the striated bottom surface of the sandstone layer. Flute casts range in size from a few millimeters in width and about a centimeter in length up to more than 10 cm by 40 cm. The place of maximum relief is usually a short distance from the rounded "beak," which rises steeply from the flat undersurface of the sandstone. Groove-cast lineation comes up abruptly to the flute casts without deviation on the front or beak end and shows only slight flaring away from the margins. Most flute casts display no lineations on their forward ends, which instead are characterized by bulbous or hummocky relief, but a quarter of the way back many exhibit clear groove-cast lineation at an angle of up to 30° from the average trend. Flute casts may line up perfectly with the prevailing lineation, but more commonly they are arranged obliquely at an angle up to 20°. In cross section most flute casts are asymmetric, and one side may even meet the bed undersurface with an overhang. Steeper flanks show several subsidiary flute casts or a region of bulbous "heads." Where several lie on the same undersurface they commonly are arranged with a "staggered" pattern: flute casts situated downstream along the lineation are found between two upstream casts. Many casts have failed to develop beaked heads, and form marked ridges subparallel to the lineation.

The characteristic shape and occurrence of flute casts associates them with similar structures already described (Clarke, 1918; Häntzschel, 1935, p. 655, fig. 5; Shrock, 1948, p. 230; Rich, 1950, p. 725). Flutings were scoured out of indurated mud by a fairly strong current

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PLATE 2.—CURRENT STRUCTURES

FIGURE 1.—OUTCROP SHOWING UNDERSIDES OF BEDS WITH THREE TYPES OF LINEATION

A: Charcoal-fragment lineation. B: Groove-cast lineation with poorly-developed flute casts (fc) and worm-burrow fillings (bf). C: Fine-grained load-cast lineation. Arrow shows current sense. Scale in centimeters. La Mortivue section, station L.

FIGURE 2.—CHARCOAL-FRAGMENT LINEATION

Detail of outcrop shown in Figure 1. Arrow shows current sense. Scale in centimeters.

FIGURE 3.—GROOVE-CAST LINEATION WITH FLUTE CASTS

Exposed on base of layer from La Mortivue section, station A-4. *Note* beaked head of large flute cast at left with bulbous subsidiary casts. Smaller flute casts lie at lower right. *Note* burrow-fillings (bf). Arrow shows current sense. Scale in centimeters.

FIGURE 4.—TOROSE LOAD CASTS

Exposed on base of bed in La Mortivue section, station A-5. S: Spiral termination. Arrow shows current sense. Scale in centimeters.

carrying sand as a cutting agent. The casts of these depressions or flutes are now preserved on the undersurfaces of many sandstone beds, hence the name flute cast. Although the term flow marking has been used for this kind of structure (Rich, 1950, p. 722; Kuenen, 1953a, p. 1051), the present writer finds it too inclusive; it seems to take in lineations and other structures as well. Flute casts are useful in giving information on the direction of flow along the lineation—the beaked or steep end is upstream. Some seem to have been localized by small depressions of probable organic origin (siphon holes?), but most no longer reveal a reason for their location. Flute casts set at a slight angle to the lineation, or with oblique lineation upon them, seem more apt to trend to the right than the left as one looks downcurrent, which suggests that the Coriolis parameter somehow played a part in their orientation.

Torose load casts.—The undersurfaces of many sandstone layers are covered with elongate ridges which pinch and swell along their trends (Pl. 2, fig. 4). Many terminate in bulbs or tear drops and some with spirals. Individuals may have nearly any shape, not necessarily elongate, but an undersurface arrayed with many forms commonly displays a marked linearity. The load casts may be 40 or 50 cm long, but most are only 10 or 15 cm in length, and a centimeter or so in width. In contrast to flute casts most torose load casts are deep at their downcurrent ends instead of at their upcurrent ends. (The sense of current flow is here definitely known from current stratification and other current structures in the same and near-by sandstone layers.) Spiral terminations (Pl. 2, fig. 4, "S"), which include the parts with highest relief, have been noticed only on the torose surfaces, and often contain the coarsest grains in view. The undersurface of the rock is streaked with a faint, fine lineation and, where several straight lines can be crowded into a millimeter, is best preserved on the flat surfaces between the torose forms, though at places it affects the upstream ends of load casts as well. The very fine texture and association with the torose casts distinguish this lineation from both groove-cast and load-cast lineation.

The load-cast origin for these forms by sinking into soft mud seems reasonable, although deformed laminae within underlying shales have not been observed. The pattern of a torose surface suggests that lower threads of a flow sank down into plastic mud which then resisted being carried away. In places deep eddies left spiral terminations. The structures must have been formed by the same flow which emplaced the overlying sandstone layer, otherwise the steep walls in the soft mud would have lost their support and the delicate sculpturings would have collapsed. Moreover, some scouring probably took place in the deeper holes so there may be transitions between load casts and flute casts.

Rücklin (1938) investigated similar forms from the Lower Muschelkalk of the Saar region, but his differ significantly from those dealt with here. His simple type (einfache Zapfen-Wülste) looks like a flute cast, but there is no associated lineation and most examples are much more irregular. The undersurface of beds is embossed with flat casts (Flachzapfen) arranged with a distinct diamond pattern (Rautenskulptur). Rücklin recognized a genetic series beginning with simple small forms and culminating in trains of plug-like welts (Zapfenzug) and large ridges at least 20 cm long (Hauptwülste). Midway in his series he placed spiral or corkscrew forms (Korkzieher-Zapfen) which are very much like the spiral terminations described here. By analogy with forms of similar shape, which were cut in indurated mud in his laboratory by running water laden with a few sand grains, Rücklin gave to the Muschelkalk "zapfen" a similar origin. He shows clearly from what direction the current came before the sculptured surface was covered by sand. His structures are therefore closely related to grooves and flutes, a conclusion supported by the present writer based upon field examination of the Muschelkalk occurrences. Rücklin's forms are different in origin from the torose load casts described here; they were cut by flow in a sense opposite to that which made the torose load casts by sinking into soft mud.

Current Structures inside of Beds

Internal lineation.—Many sandstone and siltstone layers possess laminations which allow a bed to break easily into thin slabs, some

spaced about a millimeter apart but more commonly a centimeter or two. Two major types of lineation have been observed on these partings in the Prealpine Flysch: one consisting of lined-up elongate grains or charcoal fragments, and the other consisting of faint streaks or warps in the bedding or lamination plane.

CLAST LINEATION: Elongate clastic grains, lying on bedding-plane partings, are often aligned parallel to the flow. This type of lineation, called here grain lineation, may be common but is easily overlooked in the field. Most sand grains in the rocks observed, although irregular in outline, are not markedly elongated, so that careful examination under a hand lens or binocular microscope is usually necessary for identification. It is clearly exposed, however, in a quarry at the Pont de Fillinges (Oligocene Flysch of Voirons chain) east of Geneva. In addition to the grain lineation, many slabs are sprinkled with small black clasts of charcoal and wood and plant fragments which stand out conspicuously because of their large size and dark color. Preferred orientation of these dark flecks results in charcoal-fragment lineation, which, in the examples noted, was also arranged parallel to flow (Pl. 2, fig. 2).

PARTING LINEATION: Some bedding-plane partings show faint streaks parallel to flow which are most conspicuous under oblique lighting (Pl. 1, fig. 3). These streaks are very small warps, creases, or faint shallow grooves in the bedding plane, and seem to have been formed during sedimentation and are not lines of joint intersections. Slabs with closely spaced partings in many places break steplike from parting to parting; here the orientation of the steps may follow the parting lineation. Additional work should indicate what element of the rock fabric underlies this type of lineation.

Current stratification.—Many sandstone layers in the Prealpine Flysch contain small-scale cross-stratification which is very useful in determining the direction of current flow. This cross-stratification[1], in units from one-half to 3 cm thick, may occur only at the top of beds or throughout the entire thickness in others. The most distinct is found near the base where grains are coarser and cross-stratified units thicker. In the upper parts of many layers it is much deformed, so that cross-stratifications are oversteepened and may stand nearly at right angles to the principal bedding. Several cross-stratified fragments were torn from their neighbors and rolled over while still soft and before overlying mud was laid down. Current readings must therefore be discarded from these deformed layers.

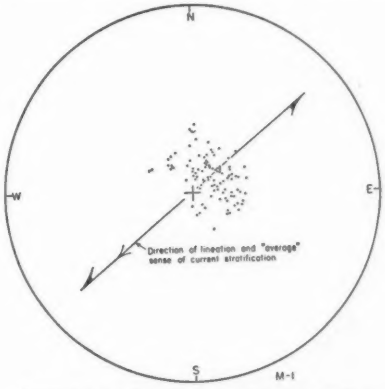


FIGURE 5.—STEREONET PLOT OF CROSS-STRATIFICATION POLES IN CURRENT STRATIFICATION

Specimen M-1, La Mortivue section; poles plotted on lower hemisphere

In the field current stratification may give approximate current direction readily, but it is often time-consuming to determine this direction precisely because individual cross-stratifications may depart in dip from the average by many degrees (Figs. 5, 6, 7). Only good exposures which afford a three-dimensional view can offer a satisfactory value. Sections through the current stratification parallel to the direction of flow display the steepest cross-stratified units, and most are inclined downcurrent. Sections normal to the current direction, on the other hand, show the cross-stratifications inclined in both directions, and appear much more irregular. Illustrations and descriptions of current bedding are included in Shrock (1948, p. 242 ff.), Pettijohn (1949, p. 127–134), and Kuenen (1953a).

Even though field studies showed that the current lineation was lined up parallel to the current direction as determined from current

[1] Terminology recommended by McKee and Wier (1953) and Ingram (1954).

stratification, two large oriented rock specimens were studied in detail in the laboratory in order to confirm this relationship. Each specimen displayed good current stratification with distinct fine-grained load-cast lineation although in each the current stratification was deformed in the upper, fine-grained part of the layer. The specimens were sawed into blocks 3 cm on a side and the surfaces polished slightly, so that when wet, the cross-stratification stood out clearly. At each corner of each block the cross-stratifications were marked carefully in pencil under a low-power (5×) binocular microscope, and the dip component of the same cross-stratification on two surfaces measured by means of a graduated turntable constructed for the purpose. Dip components were then plotted in stereographic projection, and the strike and dip of the cross-stratification obtained in the usual manner with the stereonet. The poles of these cross-stratifications on the lower hemisphere are shown in Figures 5, 6, and 7. In selecting cross-stratifications for measurement, steep and prominent ones were purposely chosen and deformed and unusual-appearing ones discarded; the diagrams therefore contain an inherent but defensible subjective bias.

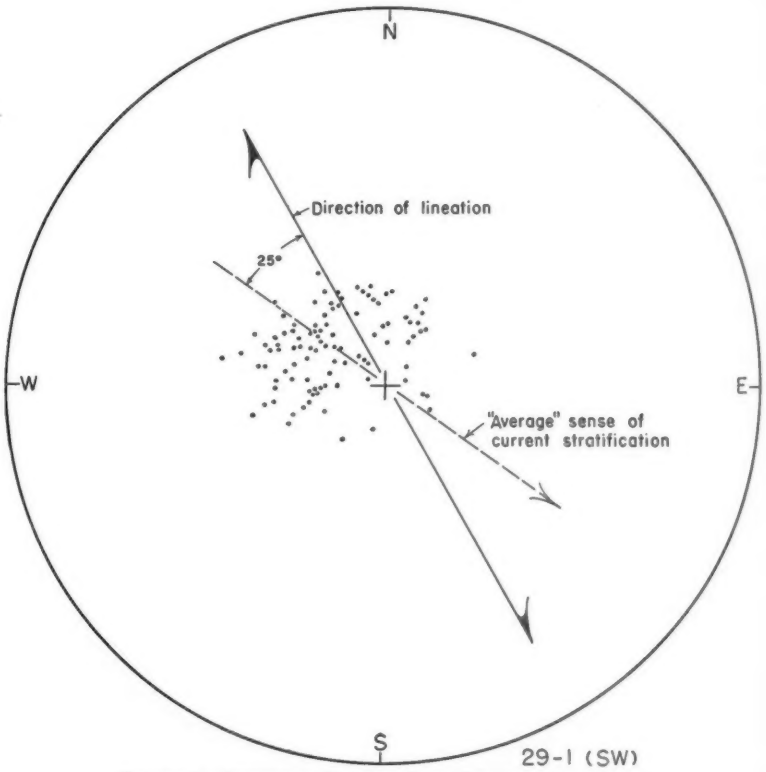


FIGURE 6.—STEREONET PLOT OF CROSS-STRATIFICATION POLES
Specimen 29-1, southwestern part, Falli-Hölli section

The specimen from locality M-1 (La Mortivue Section), 4.3 cm thick and 6 by 9 in other dimensions, yielded 100 strikes and dips of cross-stratification (200 dip components) which are plotted here in Figure 5. From specimen 29-1 (Falli-Hölli Section), 3 cm thick and 9 by 24 otherwise, 200 strikes and dips were obtained (Figs. 6, 7). The 100 from the southwest part of the block have been plotted separately from the

100 in the adjacent part of the block to the northeast as a check on the reproducibility of the same pattern from different portions of the same specimen. The direction of the lineation in 29-1, and the load-cast lineation somewhat more distinct and regular. In addition, the current stratification extends to the base of the sandstone layer, and into individual load casts

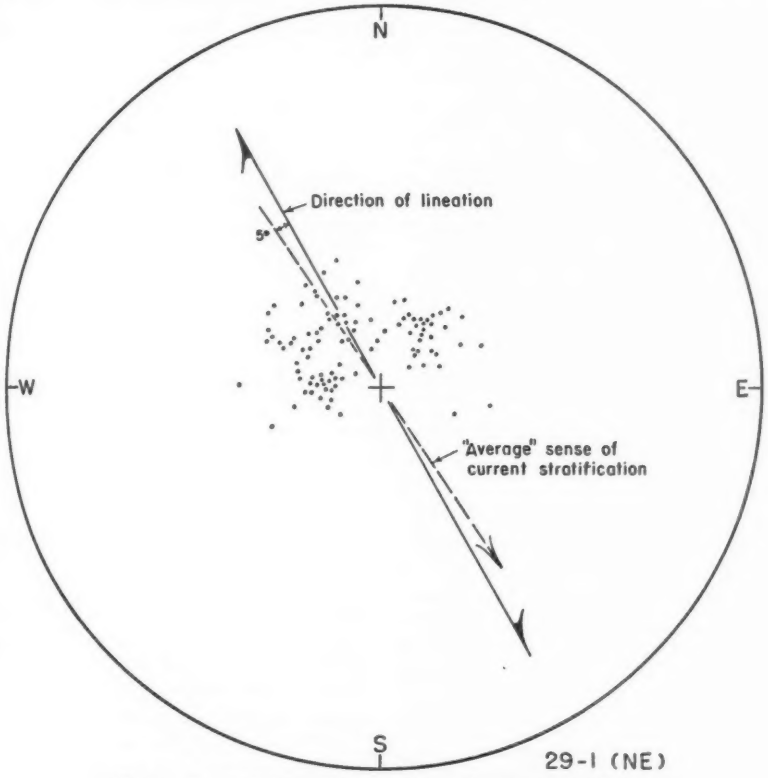


FIGURE 7.—STEREONET PLOT OF CROSS-STRATIFICATION POLES
Specimen 29-1, northeastern part, Falli-Hölli section

appearing on the bottom of the specimen is shown in each plot along with a dashed line dividing the number of poles evenly into two sectors. This dashed line, with an arrowhead, is given as an approximation of the average direction of inclination of the cross-stratifications. In view of the scatter of the points on the three diagrams, and the purposes of the study, it has not seemed worthwhile to compute the average direction precisely.

In M-1 the direction of lineation and the average direction of current bedding as determined in this fashion are the same (Fig. 5). The current bedding is more regular in M-1 than which are part of the lineation. This shows clearly that the current stratification and the lineation were formed by the *same* current, a conclusion also suggested by the diagram, so that the coincidence of the two directions may not be entirely accidental. The plots for 29-1, on the other hand (Figs. 6, 7), do not show such a coincidence of lineation and current-stratification directions and, in fact, this relation is not even improved if all 200 points are plotted on the same diagram. Although no explanation for the differences of 5 and 25 degrees between these directions seems convincing, it should be noted that the load-cast lineation on 29-1 is

somewhat braided and not nearly as regular as on M-1. Perhaps currents in 29-1 moved more sluggishly and weaved slightly from side to side.

It is not clear why poles on the diagrams are scattered so uniformly rather than clustered near the lineation direction and sparser at a distance. In fact, such an ideal pattern does not result even if all 200 points from 29-1 are plotted together. Despite these and other factors which are not as yet understood, the diagrams clearly show that there is a relationship between the lineation direction and the orientation of the cross-stratifications. Again this is proof that the two structures were formed by the same current or by two currents going in the same or nearly the same direction. Furthermore, for paleogeographic reconstructions, the diagrams indicate that it is reasonable to obtain the current direction in the field by measuring the orientation of the lineation and the sense of this direction by a quick look at the current stratification.

Convolute bedding.—Some sandstone beds, of uniform thickness over the extent of an exposure, contain complexly folded laminae with much thickening and thinning. This type of structure, termed convolute bedding by Kuenen (1952; 1953a, p. 1056), appears to have been formed as the layer was laid down inasmuch as deformed laminae, if imagined as straightened out, have a length far greater than the length of the layer in which they occur. In addition, the convoluted folds at the top commonly grade through gentler and gentler folds into relatively undeformed bedding planes, and in places the convoluted portion passes gradationally upwards into current stratification. Moreover, there is no thinning of the layer at one place and piling-up at another which would surely be the case if these were slump structures. Where the axes of the convoluted folds are regular enough for measurement they may trend normal to the current direction in the associated beds.

Ripple Marks on Top of Beds

The tops of a few sandstone layers show well-developed current ripple marks, but the structure is not common because the upper parts of many beds grade gradually into silt or clay, and these are grain sizes in which ripple marks are not apt to form. At places ripple marks make a clean-cut contrast at the top of the sandstone layer. Here the orientation of crests and troughs of ripples can be measured easily in the field and is perpendicular to associated lineations.

Field Measurements of Directional-Current Structures

One of the purposes of the present study, in addition to classifying and attempting to learn something of the origin of the current structures, was to see whether many observations in one rock sequence were consistent in direction. If the observations taken as a whole gave a significantly uniform direction, then perhaps conclusions might be drawn concerning the bottom-slope of the sea at the time of deposition, and from what direction the sediment must have come. Furthermore, consistency in direction must be found where exposures are complete or the technique will be useless where exposures are sparse.

With respect to making field observations, the rake or pitch (the angle between the line of strike and the lineation in the plane of the bedding) of the structure was measured rather than plunge because the problem of rotating the observations back to horizontal is much simpler with pitch than plunge. If the measurements are referred to bedding planes, all data can be given in two dimensions rather than three.

The directional measurements made in the field, which total 401, are listed in Table 3 and plotted as current roses in Figures 8 through 12. The pitch of the linear feature is given as measured, but rotated bearings in columns 6 and 7 may depart in some cases by one or two degrees from the correct value. This was done so that the information could be plotted conveniently on the current roses without crowding too many readings at one value. Where the exposures are particularly good, as in quarries, many more measurements were obtained (Fig. 11), which emphasizes the fact that within any continuous sequence of strata the current structures are remarkably uniform in orientation. In some cases, noted in column 8 (Table 3), only the general current sense was determined from the current bedding and the precise bear-

ing has been taken from associated lineation. The arrows from these readings on the current roses would therefore imply greater accuracy than warranted so they have been shown with

of their analogy with wind roses) have been designed to summarize and to aid in the paleogeographic interpretation of the data. Lineations where so plotted extend through the

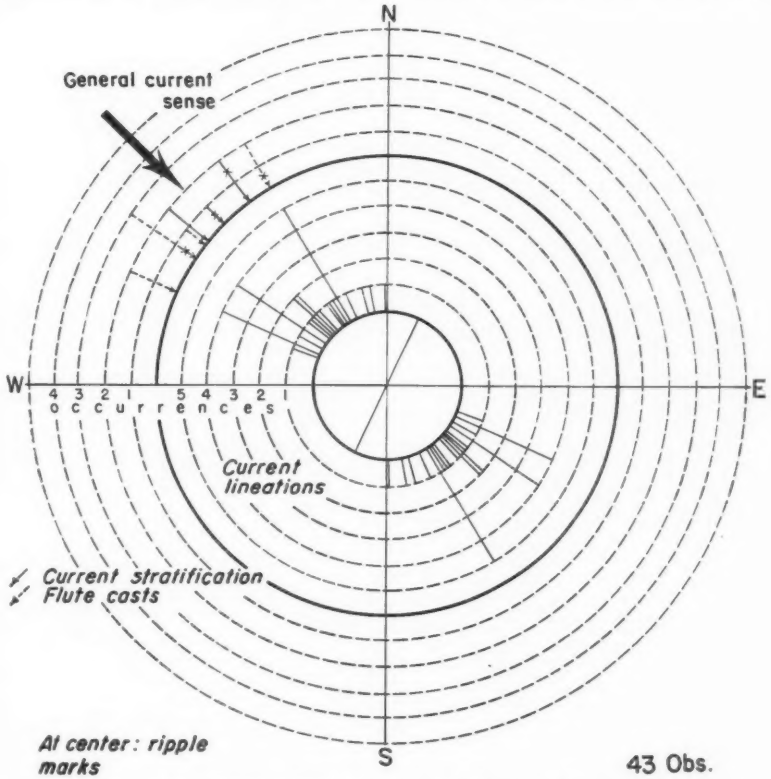


FIGURE 8.—CURRENT-ROSE DIAGRAM, LA MORTIVUE SECTION I

Stations AA to C-2. Bearings of axes of ripple marks are plotted in center; bearings of lineations in intermediate circles; and bearings of flute casts and current senses from current stratification around the outside. Observations plotted: 29 lineations, 7 current stratifications, 6 flute casts, 1 ripple mark; total: 43. *Note* that the bearing plotted for four of the current-stratification observations (shown with *x*'s) has been taken from the associated lineation (Table 3).

crosses through their stems. Since much of the classification and description was undertaken in the laboratory by using oriented specimens after completion of the field work, some of the terms in the table include more than one type. In the field not all the different kinds of substratal lineation were differentiated, although it may now be easy to do so. A summary of the number of observations of each type and for each section is included as Table 2.

The current-rose diagrams (so called because

center of the plot and into two quadrants; current stratifications and flute casts, on the other hand, are confined to only one quadrant. Care has been taken to plot together only related observations; for example, the data for Figures 8 and 9 were plotted on separate current roses because there was an abrupt change in current direction between observation C2 and C3. The data appearing on Figure 10 were plotted separately because the observations were made some distance downstream (Fig. 2).

The current roses reveal a close grouping of observations around preferred directions, which suggests that the measurements have real significance in showing current direction and sense. The heavy arrows are labelled "general" rather than "average" current sense because they have been located by inspection and do not represent arithmetical averages. Data are clustered close enough on most diagrams to suggest only one general current sense except for Figure 9 which implies two. Here the readings suggest that two sources contributed from time to time, one from the northwest and the other from the northeast, but this conclusion is not based on many observations.

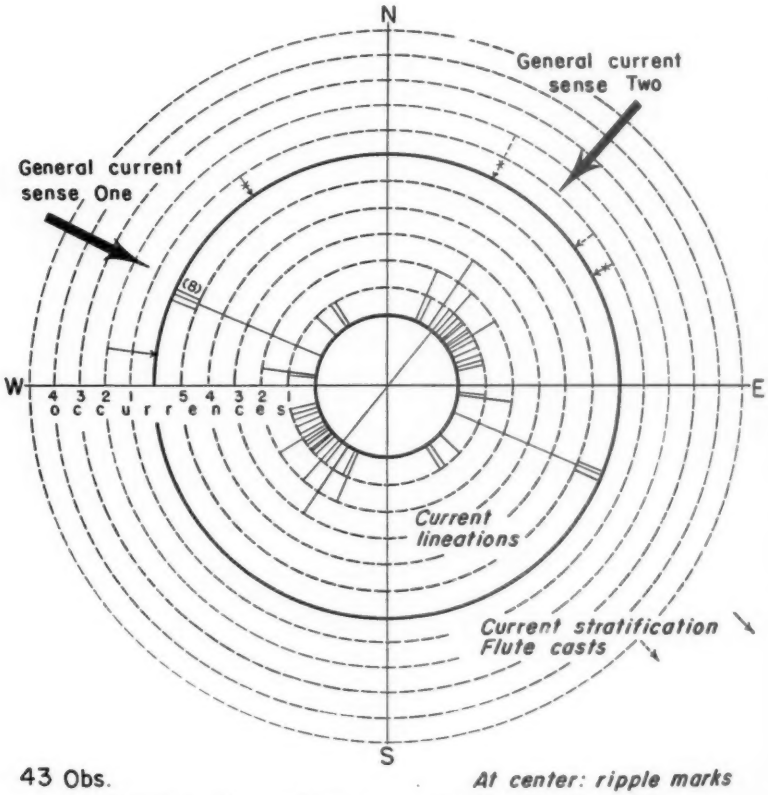


FIGURE 9.—CURRENT-ROSE DIAGRAM, LA MORTIVUE SECTION II

Stations C-3 to I. Observations plotted: 35 lineations, 5 current stratifications, 2 flute casts, 1 ripple mark; total 43. *Note* that the bearing plotted for three of the current stratification observations (shown with x's) has been taken from the associated lineations (Table 3).

SITE AND MANNER OF DEPOSITION

A statement on the site and manner of deposition of the Gurnigel Flysch must take into account other features besides the current structures described here. Thin sandstone beds, graded or partly graded, contrast markedly with interbedded fine-grained shales, mudstones, or marls. These contrasting layers are persistent over remarkable distances in relation to their thinness and retain intricate textural and structural details. Many pelitic sediments display thin laminae, some crinkled but unbroken, and sandstones contain elaborate features which are preserved and found in out-

crops. Unless sedimentation took place in relatively quiet or sheltered water, these details would have been destroyed by reworking incited by waves, winds, or currents. It therefore seems likely that sedimentation took place in water deeper than "wave base" and further that the bottom water was homogeneous and confined to a closed basin so that most currents were suppressed, including those caused by tides, internal waves, and density layers in the water. Although benthonic organisms lived in such an environment, as shown by the numerous trails and burrows, they seldom succeeded in reworking the sediments so that structural details were destroyed.

The strata dealt with here display most of the features purported to characterize sediments laid down by turbidity currents (Kuenen and Carozzi, 1953, p. 364). Among those recognized in the Gurnigel Flysch are the interstratification of fine-grained deposits with coarser-grained beds, regular bedding, slump structures, and pull-aparts. The sandstone layers show graded bedding, poor sorting at each horizon, sharp lower contacts, load casts, convolute bedding, shale inclusions, flow marks, current ripple marks, and small-scale current stratification. In addition, the thicker sandstone beds have the coarser grains and each bed tends to maintain its own characteristic features, such as

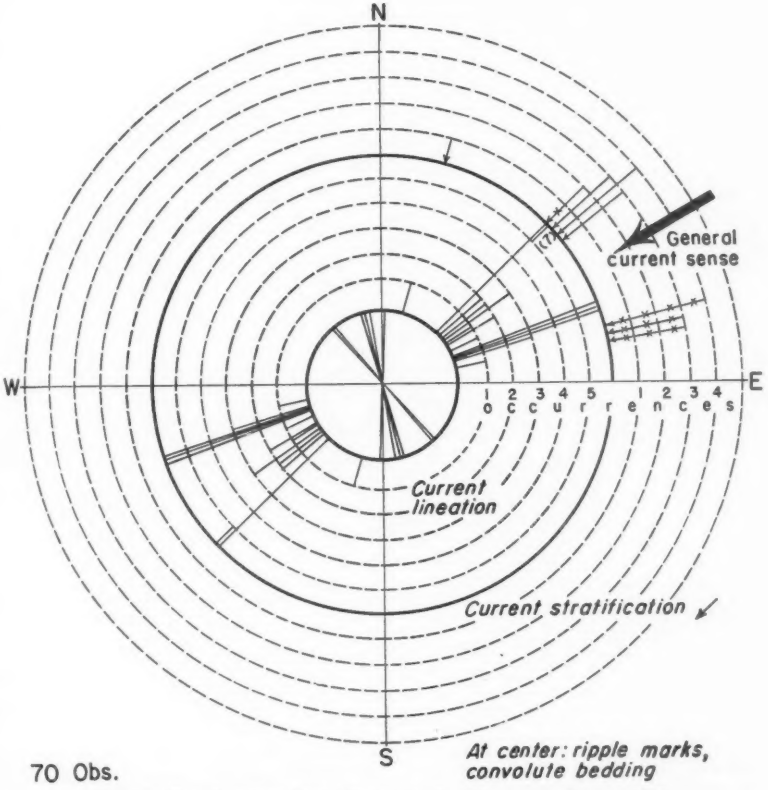


FIGURE 10.—CURRENT-ROSE DIAGRAM, LA MORTIVUE SECTION III

Stations K to M. Observations plotted: 41 lineations, 23 current stratifications, 6 ripple marks and convolute beddings; total: 70. *Note* that the bearing plotted for 11 of the current stratification observations (shown with x's) has been taken from the associated lineation (Table 3).

convolute bedding, load casts, and grading, over its length of exposure. We therefore conclude that the sandstone layers of the Gurnigel Flysch were laid down by turbidity currents, a coarser part may pinch and swell markedly, so that at one place it has nearly disappeared and at another thickened to about half a centimeter. In places the thicker parts clearly show

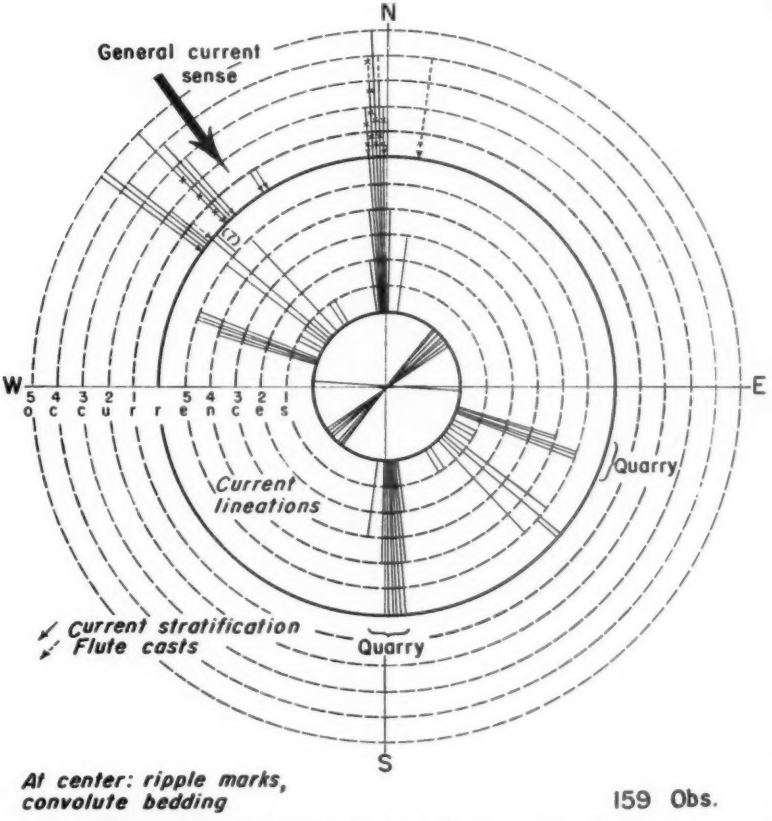


FIGURE 11.—CURRENT-ROSE DIAGRAM, SCHWARZENBÜHL SECTION

Observations plotted: 95 lineations, 49 current beddings, 6 flute casts, 10 ripple marks and convolute beddings; total: 159. *Note* that many of the observations came from two quarries, and that the bearing plotted for 15 of the current-stratification observations (shown with x's) has been taken from the associated lineation (Table 3).

conclusion already reached for similar rocks in the Oligocene Flysch of the Voirons chain east of Geneva (Kuenen and Carozzi, 1953, p. 366).

Although it seems clear that the sandstone beds were deposited by turbidity currents, the pelitic layers between were not necessarily deposited in the same fashion. In many places shales are laminated into graded units as thin as 2 mm. These laminae are commonly continuous across an outcrop, but the lower current stratification, and the shape of the pinches and swells often suggests the profile of current ripples. It therefore seems likely that much of the muddy material was also laid down by small, slow, and dilute turbidity currents, which may have originated frequently in comparison to the more powerful ones responsible for the sandstone beds. Mudstone interlayers, which under the hand lens exhibit no laminae, grading, or structure whatsoever, may also have

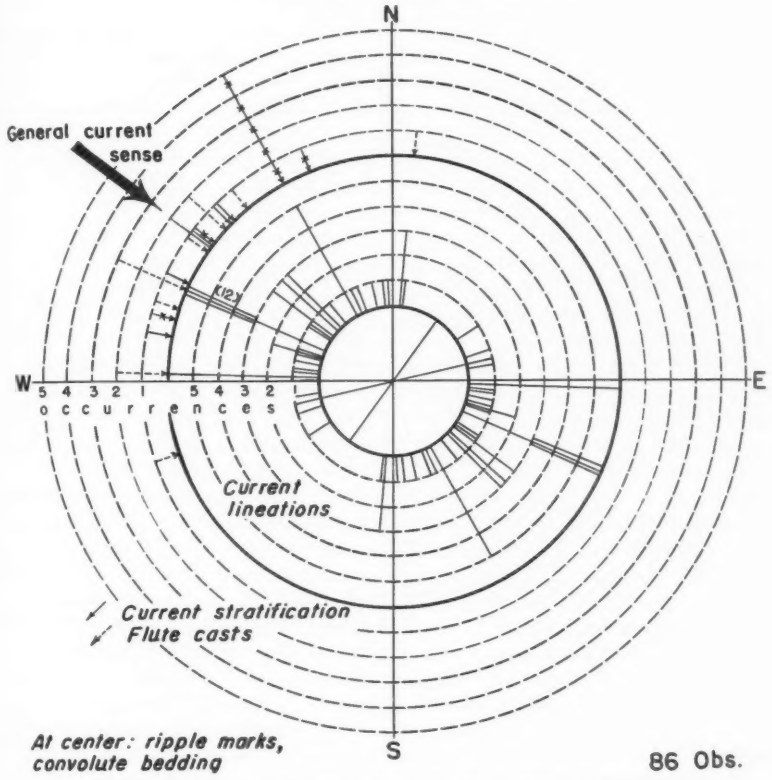


FIGURE 12.—CURRENT-ROSE DIAGRAM, FALLI-HÖLLI SECTION

Observations plotted: 57 lineations, 15 current stratifications, 12 flute casts, 1 ripple mark, and 1 convolute bedding; total: 86. *Note* that the bearings plotted for 8 of the current-stratification observations (shown with x's) has been taken from the associated lineation (Table 3).

TABLE 2.—NUMBER OF DIRECTIONAL OBSERVATIONS OF EACH TYPE

	La Mortivue	Schwarzenbühl	Falli-Hölli	Total
Substratal lineation (undifferentiated)	35	64	13	112
Grove-cast lineation	41	13	36	90
Load-cast lineation	0	8	5	13
Flute cast	8	6	12	26
Charcoal-fragment lineation	5	5	0	10
Parting lineation	24	5	3	32
Current stratification	35	49	15	99
Convolute bedding	1	8	1	10
Ripple mark	7	1	1	9
Totals	156	159	86	401

been deposited by turbidity currents which had available to them only homogeneous fine-grained sediment. It is equally possible, however, that the mudstones accumulated by slow settling of dispersed sediment from surface layers of the sea. Some fine sediment may have

been contributed to the basin by surface inflow (hypopycnal inflow) whereas coarser sediments accumulated inshore and then moved basinward in turbidity currents (hyperpycnal inflow) (Bates, 1953, p. 2125).

In the quiet of the relatively deep basin of sedimentation, the arrival of a turbidity current was probably the only event to contrast with the gentle rain of sediment from above and the occasional stirring of organisms. As each turbidity current roiled into the basin it would of course flow down the slope and, losing speed and turbulence, deposit sediment in the lowest place within reach. Successive flows coming from the same source would build a gentle fan or submarine delta grading at its distal end into a featureless plain. In such an environment, where each turbidity current seeks its own level, uniformity in the orientation of the current structures such as those in the Gurnigel Flysch seems likely. The turbid flows probably moved into the region of sedimentation as broad sheets and not as confined currents twisting and turning as they meander. It therefore seems reasonable that measurements of current structures in such strata, where we have enough to deal with averages, can indicate the slope direction of the bottom. In turn, this direction gives information on the direction to the source of the sediment.

Studies to date suggest that much of the sediment travelling into deep near-shore basins by way of turbidity currents goes through submarine canyons (Shepard, 1951; Handin, 1951; Crowell, 1952; Dill, *et al.* 1954). Under these circumstances the mouths of the canyons scattered along a coast would act as point-sources for the turbid flows arriving in the deeper basin. Current structures in rocks laid down some distance from shore would point back upslope along radii of the submarine fans or deltas toward the mouths of these submarine canyons, and not directly toward nearest land. Within the "flank-fan depressions" near shore and between canyons, but in the deep basin, current structures might even show orientation parallel to the shore. Thus they indicate the source only approximately, and the Gurnigel Flysch can be said to have its source somewhere to the northwest. Two point-sources (canyons ?) may have contributed to rocks in the La Mortivue section, inasmuch as there is an overlapping of two distinct current directions in the middle of the section (Fig. 9).

Data are insufficient to relate observations on the Gurnigel Flysch to more precise conditions in the basin of deposition, such as depth and actual slope of the basin floor. In other regions, where outcrops are better and the structure is less complex, it may be possible to obtain much more information from the study of these current structures, particularly in combination with other approaches. For example, independent information on thickening and thinning of coarse members and on clastic ratios would be very helpful.

The Wildflysch associated with the Gurnigel sandstone, with its conglomerate, breccia, coarse sandstone, and exotic blocks, contains many of the same structures accredited to turbidity currents, and possibly much of this series was laid down by them as well. No doubt the flows were larger or more powerful in order to bring in coarse debris, and some of the deposits of unsorted rubble may have accumulated as slides or slumps rather than from true turbidity currents. Perhaps the land area was more youthful and rugged during and immediately following an orogenic impulse so that coarser debris was contributed to the sea, or submarine slopes may have been steeper. It seems possible that as the Wildflysch accumulated, the source region became less rugged and the basin filled, so that gradients were gentler when the overlying Gurnigel Flysch was laid down.

The environment proposed for the accumulation of the Flysch of the External Prealps is probably similar to that off the coast of southern California today (Tercier, 1947, p. 180). Here deep basins lie between submarine ridges, some of which culminate with islands (Shepard and Emery, 1941). The floors of near-shore basins are underlain by sand and mud which are accumulating today (Revelle and Shepard, 1939; Shepard, 1951), and evidence suggests that turbidity currents moving down submarine canyons are playing a significant role (Shepard, 1951; Crowell, 1952). Late Tertiary deposits now exposed on the near-by continent evidently accumulated in a similar environment and are also very much like the Prealpine Flysch. Pliocene rocks of the Ventura Basin, for ex-

ample, contain many of the same structures dealt with here and are known by Foraminifera ecology to have accumulated in deep water (Natland and Kuenen, 1951). Elsewhere in southern California other Tertiary formations were also clearly laid down by turbidity currents (Kuenen, 1953a; Winterer and Durham, in press). In northern California the extremely thick series of Jurassic and Cretaceous rocks (Knoxville, Shasta, Chico) along the western side of the Sacramento Valley was deposited mainly by this process. Rocks with similar structures, and so perhaps formed in the same way, have been reported from many other parts of the world (Clarke, 1918; Gürich, 1933; Häntzschel, 1935; Kieslinger, 1937; Kuenen, 1948; Shrock, 1948, p. 131; Kuenen and Carozzi, 1953).

In summary, the Gurnigel Flysch evidently accumulated in a relatively deep marine basin not far from a landmass or island chain to the northwest. From time to time sediment came to rest on the floor of the basin through the action of turbidity currents which came from sources, perhaps the mouths of submarine canyons, along the submarine base of the landmass.

Tectonics of External Prealps

The external Prealps consist of a complex folded and faulted mass of Mesozoic and Tertiary rocks within the giant Prealpine klippe, a far-travelled mass. The geometry of structures, in combination with stratigraphic studies, shows that rocks in the klippe have moved northwest from their original place of accumulation far to the southeast (Figs. 13, 15). An understanding of the history of the Prealps has resulted largely from efforts to reconstruct the original distribution of stratigraphic units, often from very scanty and broken evidence. The fact that the Gurnigel Flysch in the External Prealps was laid down in a marine trough with a land source lying to the northwest

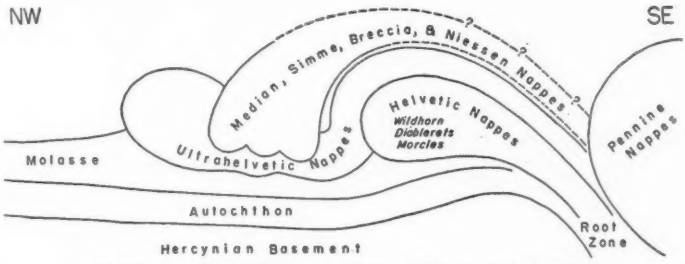


Figure 13.—Scheme of Tectonic Units in the Prealps

(Fig. 14) must be considered in a reconstruction of Paleocene-Eocene paleogeography.

The general structure of the Prealps is shown diagrammatically in Figure 15; specific information is given by Jeannet (1922), Heim (1922), Lugeon and Gagnebin (1941), Tercier (1952), and Cadisch (1953, p. 172–179, 182–185, 196–217, 311–319). Summaries in English, now considerably outdated, appear in Bailey (1935, p. 57–89) and Collet (1935, p. 233–254). The Prealps consist of at least six or eight superimposed complex thrust-plates or nappes, frequently involuted, and folded and faulted after emplacement so that locally complicated structures have resulted. Some nappes, such as the Breccia and the Simme, began to develop in the Cretaceous—debris eroded from them is incorporated in Cretaceous Flysch (Campana, 1943). The growth of these and other nappes continued slowly through the Eocene and then culminated with major paroxysms in the Oligocene and Pliocene when the nappe complex reached its present structure and position. The Ultrahelvetic nappes, although actually initiated later than the Breccia and Simme nappes, grew rapidly in front of them (northwest) and reached a position far to the northwest of the Ultrahelvetic root zone and now underlie the other Prealpine nappes. As the upper nappes rode over the underlying complex an Ultrahelvetic mass separated and moved even farther to the northwest, and now makes

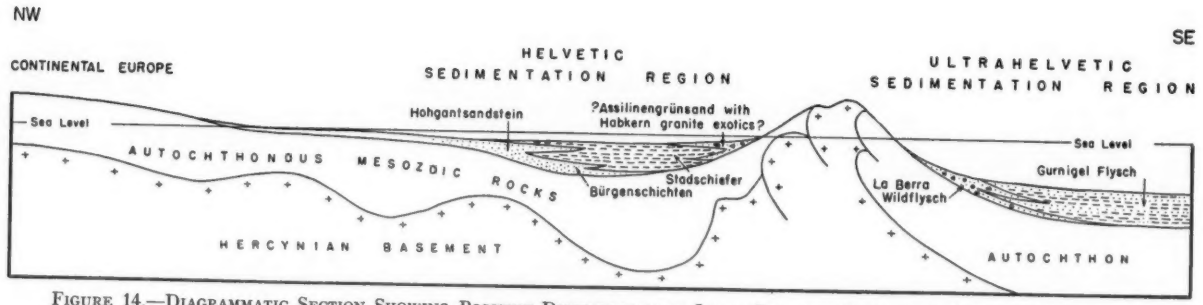
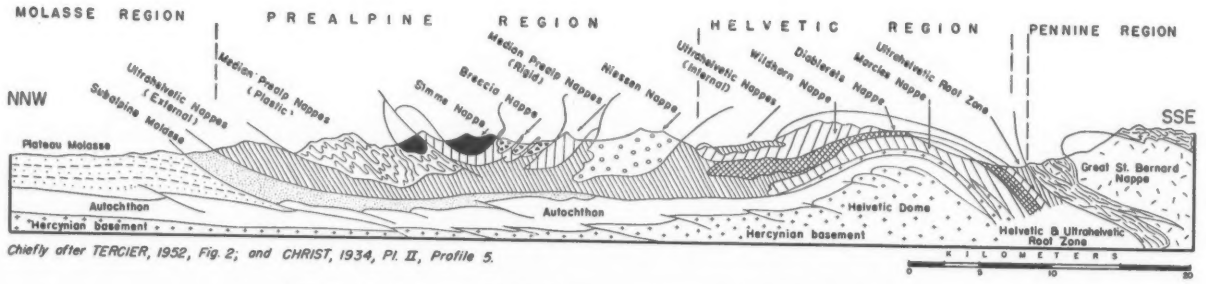


FIGURE 14.—DIAGRAMMATIC SECTION SHOWING POSSIBLE DISTRIBUTION OF LAND, SEA, AND REGIONS OF SEDIMENTATION DURING THE PALEOCENE AND EOCENE



Chiefly after TERCIER, 1952, Fig. 2; and CHRIST, 1934, Pl. II, Profile 5.

FIGURE 15.—TECTONIC CROSS SECTION THROUGH THE ROMANDE PREALPS

up the External Prealps. Shreds of these lower nappes left behind allow Ultrahelvetic rocks to be followed southeastward back over and beyond the Helvetic nappes in the High Calcareous Alps and into their probable root zone along the Rhone River, the Sion-Sierre zone (Lugeon and Gagnebin, 1941; Badoux, 1946). Rocks of the External Prealps can therefore be traced back to the vicinity of their probable root-zone at least 56 km (35 miles) to the southeast.

If this structure is "unrolled," Helvetic stratigraphic units seem to grade into Ultrahelvetic units still farther to the southeast, although the latter are now found to the northwest (Heim, 1922; Tercier, 1952). An orderly reconstruction of the original Mesozoic stratigraphy can be worked out, which shows thickening, with some facies changes, from the thin autochthonous sequence overlying the central massifs (Helvetic dome) into the Helvetic region to the southeast, and then in turn into the Ultrahelvetic region. This reconstruction follows reasonably from comparisons of the Triassic, Jurassic, and Lower Cretaceous rocks, but younger units (including the Flysch and Wildflysch) change facies too rapidly and appear too different in different nappes to help much in working out the paleogeography. Understanding of the conditions during the late Cretaceous and early Tertiary is therefore much more dependent on other techniques, such as information on source areas from studies on clast types. Perhaps the data from current structures described here will help clarify Paleocene-Eocene paleogeography when the major deformation of the Alpine orogeny was beginning to gather momentum.

Clasts from the Eocene Wildflysch in the La Berra region (just southwest of the Falli-Hölli section) have been studied in detail by Tercier (1928; 1952). This Wildflysch, which is partly equivalent and only slightly older (Upper Lutetian) than the Gurnigel sandstone, contains clasts in conglomerate and breccia beds and as scattered exotics and also displays some current structures suggesting derivation from the north and northwest. Crystalline clasts identified by Tercier include 18 types of porphyry, 8 types of granite, a diorite, and 6 schists and gneisses. Many of the clasts of sedimentary rocks contain fossils or other distinctive characteristics which allow them to be identified stratigraphically. These clasts, which are mostly limestones with some sandstones, quartzites, and radiolarities, range in age from Permo-Carboniferous to late Cretaceous. Tercier concludes that the La Berra Wildflysch was derived from an area where rocks of the Ultrahelvetic facies were exposed, and is quite certain that the central massif (Helvetic dome) did *not* contribute. In addition, some of the crystalline clasts in the Jurassic rocks came from the same basement area which shed debris to the Wildflysch. The large amount of detrital limestone in the Gurnigel sandstone presumably has been reworked from older Ultrahelvetic limestones.

The affinity of some of the clasts to the granitic rocks of the Err-Bernina region of eastern Switzerland would suggest that these were derived from an Austride or Lower East Alpine source, but this correlation has not been universally accepted (Tercier, 1952; Cadisch, 1953, p. 183). According to Tercier (1952, p. 37) it seems most likely that the granitic and metamorphic clasts came from a crystalline basement beneath the Ultrahelvetic facies.

In the reconstruction of the Eocene setting of the Gurnigel sandstone the distribution of near-by related Paleocene-Eocene units is significant. In the Helvetic sedimentation region, which originally lay just to the northwest of the Ultrahelvetic area, a distinct Eocene transgression toward the northwest is discernible. Arnold Heim has carefully "unrolled" the Helvetic nappes and shows a reconstruction of the Helvetic Eocene (1922, fig. 97). At the northwestern border of the Helvetic area, Auversian sandstone (Hohgantsandstein) grades southeastward into shale (Stadschiefer) which in turn grades downward into Lutetian sandstones (Bürgenschichten) in a manner suggesting a source at the northwest and deeper water relatively distant from shore at the southeast. With these stratigraphic relations so clearly established, any Flysch lying still further to the southeast and grading into these units must have had a source at the southeast because coarser sandstones, conglomerates, breccias, and exotic blocks could not have been transported from the northwest through seas where finer sediments were accumulating. By a

FIGURE 15.—TECTONIC CROSS SECTION THROUGH THE ROMANDE PREALPS

TABLE 3.—FIELD MEASUREMENTS OF DIRECTIONAL-CURRENT STRUCTURES

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
La Mortivue Section	AA	1	N39E	72SW	N69W		Groove-cast lineation	8
		2	60S (Ov)	75SW	N66W	N66W	Groove-cast lineation w/ up-dip flute casts	
		3		75SW	N66W		Groove-cast lineation	
		4		80SW	N61W		Groove-cast lineation	
		5		70SE	N31W	N31W	Groove-cast lineation w/ up-dip flute casts	
		6		70SE	N31W		Groove-cast lineation	
		7		70SE	N31W		Groove-cast lineation	
		8		70SE	N31W		Groove-cast lineation	
		9		85SW	N56W		Groove-cast lineation	
		10		75SW	N66W		Groove-cast lineation	
		11		75SW	N66W	N66W	Groove-cast lineation w/ flute casts	
		12		85SW	N56W	N56W	Groove-cast lineation w/ flute casts	
		13		85SW	N56W		Parting lineation	
	A	1		90		N51W	Current stratification, up dip	
		2		90		N51W	Current stratification, up dip	
		3		70SE	N31W		Groove-cast lineation	
		4		70SE		N31W	Current stratification, up dip**	
		5		76SE	N37W		Groove-cast lineation	
		6		76SE		N37W	Current stratification, up dip**	
		7		76SE		N37W	Current stratification, up dip**	
		8		50SE	N11W		Groove-cast lineation	
		9		73SE	N34W		Groove-cast lineation	
		10		83SE	N44W		Groove-cast lineation	
		11		90	N51W		Groove-cast lineation	
		12		40SE	N 1W		Groove-cast lineation	
		13		75SE	N36W		Charcoal-fragment lineation	
		14		84SE	N45W		Charcoal-fragment lineation	
		15		84SE		N45W	Current stratification, up dip**	
		16		85SE	N46W		Groove-cast lineation	
		17		85SE	N46W		Groove-cast lineation	
		18		84SE	N45W		Groove-cast lineation	
		19		85SW	N56W	N56W	Groove-cast lineation w/ flute casts	
		20		85SW		N56W	Current stratification, up dip**	
		21		64SE	N25W		Parting lineation	
	B	1	N24E 45S (Ov)		N24E		Ripple marks roughly parallel to strike	
	C	1	N18E	35NE	N17W		Groove-cast lineation	
		2	74E	70E	N52W	N52W	Groove-cast lineation w/ flute casts	
		3	(Ov)	21S	N39E		Groove-cast lineation	9
		4		5S	N23E		Groove-cast lineation	
		5		5S	N23E		Groove-cast lineation	
	D	1	N46E	15SSW	N61E		Groove-cast lineation	
		2	71SE	14SSW	N60E		Groove-cast lineation	
		3	(Ov)	14SSW		N60E	Current stratification from NE**	

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
La Mortivue Section	D	4		28SSW	N74E		Groove-cast lineation	9
		5		8SSW	N54E		Charcoal-fragment lineation (Fault zone)	
	E	1	N31E	5SSW	N26E	N26E	Groove-cast lineation w/ flute casts	
		2	79SE	5SSW		N26E	Current stratification from NE**	
		3		22NE	N53E	N53E	Groove-cast lineation w/ flute casts	
		4	N15E	18NE	N33E		Parting lineation	
		5	60E	18NE	N33E		Parting lineation	
		6		18NE	N33E		Parting lineation	
		7		18NE	N33E		Parting lineation	
		8		26NE	N41E		Groove-cast lineation	
		9		26NE	N41E		Groove-cast lineation	
	F	1	N31E 84NW	65W	N84W		Parting lineation	
	G	1	N38E	84SE	N46W		Groove-cast lineation	
		2	68SE	38S	N76E		Groove-cast lineation	
		3	(Ov)	16S	N54E		Groove-cast lineation	
		4		33S	N71E		Groove-cast lineation	
		5		60S	N82W		Groove-cast lineation	
		6			N38E		Ripple marks parallel to strike	
		7		60S	N82W		Groove-cast lineation	
		8		75S	N67W		Parting lineation	
		9		75S	N67W		Parting lineation	
		10		75S	N67W		Parting lineation	
		11		75S	N67W		Parting lineation	
		12		75S	N67W		Parting lineation	
		13		75S	N67W		Parting lineation	
		14		75S	N67W		Parting lineation	
		15		75S	N67W		Parting lineation	
		16		60S		N82W	Current stratification	
		17		60S		N82W	Current stratification	
	H	1	N25E	58SE	N33W		Substratal lineation	
		2	64SE	60SE		N35W	Current stratification from NW**	
		3		60SE	N35W		Substratal lineation	
	I	1	N20E	30S	N50E		Substratal lineation	
		2	56E	26S	N46E		Substratal lineation	
		3	(Ov)	46SSE	N66E		Substratal lineation	
	K	1	N46E	30S		N16E	Current stratification	10
		2	37SE	0		N46E	Current stratification	
		3		0	N46E		Parting lineation	
		4		0	N46E		Parting lineation	
		5		0	N46E		Parting lineation	
		6		0	N46E		Parting lineation	
		7		0	N46E		Parting lineation	
		8		0	N46E		Parting lineation	
		9		0	N46E		Parting lineation	
		10		30S	N16E		Parting lineation	

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
La Mortivue section	K	11		0		N46E	Current-stratification from NE**	10
		12		90SE	N44W		Ripple marks, irregular	
		13		45S	N 1E		Ripple marks, irregular	
		14		12NE	N58E		Parting lineation	
	L	1	N78E	↑		N76E	Current stratification from E**	
		2	77S			N76E	Current stratification from E**	
		3	(Ov)			N76E	Current stratification from E**	
		4				N76E	Current stratification from E**	
		5				N78E	Current stratification from E**	
		6				N78E	Current stratification from E**	
		7				N78E	Current stratification from E**	
		8				N80E	Current stratification from E**	
		9				N80E	Current stratification from E**	
		10		↓		N80E	Current stratification from E**	
		11		↑	N70E		Substratal lineation	
		12			N70E		Substratal lineation	
		13			N70E		Substratal lineation	
		14			N70E		Substratal lineation	
		15			N70E		Substratal lineation	
		16			N70E		Substratal lineation	
		17			N71E		Substratal lineation	
		18			N71E		Substratal lineation	
		19			N71E		Substratal lineation	
		20			N71E		Substratal lineation	
		21			N71E		Substratal lineation	
		22			N71E		Substratal lineation	
		23			N72E		Substratal lineation	
		24			N72E		Substratal lineation	
		25			N72E		Substratal lineation	
		26			N72E		Substratal lineation	
		27			N72E		Substratal lineation	
		28			N72E		Substratal lineation	
		29			N73E		Substratal lineation	
		30			N73E		Substratal lineation	
		31		↓	N73E		Substratal lineation	
		32		0	N78E		Charcoal-fragment lineation	
		33		90	N10W		Ripple marks, deformed	
		34		90	N12W		Ripple marks, deformed	
		35		90	N14W		Ripple marks, deformed	
	M	1	N50E	↑	N50E		Substratal lineation	
		2	84SE		N50E		Substratal lineation	
		3			N53E		Substratal lineation	
		4			N53E		Substratal lineation	
		5			N55E		Substratal lineation	
		6			N55E		Substratal lineation	
		7			N55E		Substratal lineation	
		8			N59E		Substratal lineation	
		9		↓	N59E		Substratal lineation	

10 measurements approximately parallel to strike

21 measurements between 5 and 8E

9 measurements between 0 and 9ENE

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
La Mortivue section	M	10		10 measurements approximately parallel to strike		N48E	Current stratification	10
		11				N48E	Current stratification	
		12				N48E	Current stratification	
		13				N50E	Current stratification	
		14				N50E	Current stratification	
		15				N50E	Current stratification	
		16				N50E	Current stratification	
		17				N52E	Current stratification	
		18				N52E	Current stratification	
		19				N52E	Current stratification	
		20		90	N40W		Convolute bedding	
		21		15ENE	N65E		Charcoal-fragment lineation	
Schwarzenbühl Section	1	1	N87E	90	N		Load-cast lineation, large	11
		2	34S	90		N 1W	Current stratification, down dip**	
		3		90		N 1W	Current stratification, down dip**	
		4		85SSE	N 8W		Groove-cast lineation	
		5		90		N 2W	Current stratification, down dip**	
		6		0	N85W		Convolute bedding	
		7		0	N89W		Ripple marks	
		8		90	N		Substratal lineation	
		9		90	N		Substratal lineation	
		10		90	N		Substratal lineation	
		11		90	N 1W		Substratal lineation	
		12		90	N 1W		Substratal lineation	
		13		90	N 1W		Substratal lineation	
		14		90	N 1W		Substratal lineation	
		15		90		N 2W	Current stratification	
		16		90		N 2W	Current stratification	
		17		90		N 2W	Current stratification	
		18		90		N 3W	Current stratification	
		19		85SSE	N 8W		Substratal lineation	
		20		39 measurements, down dip, and approximately perpendicular to strike		N 3W	Flute casts	
		21				N 3W	Flute casts	
		22			N 1W		Parting lineation	
		23			N 1W		Parting lineation	
		24			N 2W		Parting lineation	
		25			N 2W		Parting lineation	
		26	N87E		N 2W		Parting lineation	
		27	34S		N 2W		Charcoal-fragment lineation	
		28			N 2W		Charcoal-fragment lineation	
		29			N 2W		Charcoal-fragment lineation	
		30			N 3W		Charcoal-fragment lineation	
		31			N 3W		Charcoal-fragment lineation	
		32			N 3W		Substratal lineation	

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
Schwarzenbühl Section	1	33		↑	N 3W		Substratal lineation	11
		34			N 3W		Substratal lineation	
		35			N 3W		Substratal lineation	
		36			N 4W		Substratal lineation	
		37			N 4W		Substratal lineation	
		38			N 4W		Substratal lineation	
		39			N 4W		Substratal lineation	
		40			N 4W		Substratal lineation	
		41			N 4W		Substratal lineation	
		42			N 5W		Substratal lineation	
		43			N 5W		Substratal lineation	
		44			N 5W		Substratal lineation	
		45			N 5W		Substratal lineation	
		46			N 5W		Substratal lineation	
		47			N 5W		Substratal lineation	
		48			N 6W		Substratal lineation	
		49			N 6W		Substratal lineation	
		50			N 6W		Substratal lineation	
		51			N 6W		Substratal lineation	
		52				N 3W	Current stratification, down dip**	
		53				N 3W	Current stratification, down dip**	
		54				N 4W	Current stratification, down dip**	
		55				N 4W	Current stratification, down dip**	
		56				N 4W	Current stratification, down dip**	
		57				N 4W	Current stratification, down dip**	
		58				N 5W	Current stratification, down dip**	
		59		↓		N 5W	Current stratification, down dip**	
	2	1	N83E	90	N 7E	N 7E	Groove-cast lineation w/ flute casts	
		2	38S	90	N 7E	N 7E	Groove-cast lineation w/ flute casts	
		3		90	N 7E	N 7E	Groove-cast lineation w/ flute casts	
		4		90		N 7E	Current stratification, down dip**	
	3	1	N45E	90		N45W	Current stratification, down dip**	
		2	34SE	90		N45W	Current stratification, down dip**	
		3		0	N45E		Convolute bedding	
		4		0	N45E		Convolute bedding	
		5		0	N45E		Convolute bedding	
		6		0	N45E		Convolute bedding	
		7		80S	N35W		Groove-cast lineation	
		8		90		N45W	Current stratification, down dip**	
		9		90		N45W	Current stratification, down dip**	
		10		90		N45W	Current stratification, down dip**	
		11		75E	N60W		Load-cast lineation	
	4	1	N42E	90	N48W		Groove-cast lineation	
		2	30SE	45ENE	N87E		Groove-cast lineation	
		3		75E	N63W		Groove-cast lineation	
	5	1	N58E	90	N32W		Groove-cast lineation	

39 measurements, down dip, and approximately perpendicular to strike

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
Schwarzenbühl Section	5	2	42S	90		N32W	Current stratification, fair	11
	6	1	N40E	Down-dip approximately perpendicular to strike ↑	N50W		Load-cast lineation	
		2	30SE		N50W		Load-cast lineation	
		3			N50W		Load-cast lineation	
		4			N50W		Load-cast lineation	
		5			N50W		Load-cast lineation	
		6			N50W		Load-cast lineation	
		7		↓	N50W	N50W	Groove-cast lineation w/ flute casts	
		8		0	N40E		Convolute bedding	
		9		0	N40E		Convolute bedding	
		10		0	N40E		Convolute bedding	
		11		70E	N68W		Groove-cast lineation	
		12		70E	N69W		Groove-cast lineation	
		13		70E	N71W		Groove-cast lineation	
	7	1	N57E	70E	N53W		Substratal lineation	
		2	38S	70E	N53W		Substratal lineation	
		3		70E	N53W		Substratal lineation	
		4		70E	N53W		Substratal lineation	
		5		70E	N53W		Substratal lineation	
		6		90		N33W	Current stratification, fair	
	8	1	N36E	90	N54W		Substratal lineation	
		2	21SE	90		N54W	Current stratification, down dip	
	9	1	N37E	25 measurements between 70 and 75E ↑	N68W		Substratal lineation	
		2	29S		N68W		Substratal lineation	
		3			N68W		Substratal lineation	
		4			N68W		Substratal lineation	
		5			N69W		Substratal lineation	
		6			N69W		Substratal lineation	
		7			N69W		Substratal lineation	
		8			N69W		Substratal lineation	
		9			N70W		Substratal lineation	
		10			N70W		Substratal lineation	
		11			N70W		Substratal lineation	
		12			N70W		Substratal lineation	
		13			N70W		Substratal lineation	
		14			N71W		Substratal lineation	
		15			N71W		Substratal lineation	
		16			N71W		Substratal lineation	
		17			N71W		Substratal lineation	
		18			N72W		Substratal lineation	
		19			N72W		Substratal lineation	
		20			N72W		Substratal lineation	
		21			N72W		Substratal lineation	
		22			N73W		Substratal lineation	
		23			N73W		Substratal lineation	
		24			N73W		Substratal lineation	
		25		↓	N73W		Substratal lineation	

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
Schwarzenbühl Section	9	26		90		N51W	Current stratification, down dip	11
		27		15 measurements, down dip, ← approximately perpendicular → to strike		N51W	Current stratification, down dip	
		28				N52W	Current stratification, down dip	
		29				N52W	Current stratification, down dip	
		30				N52W	Current stratification, down dip	
		31				N52W	Current stratification, down dip	
		32				N53W	Current stratification, down dip	
		33				N53W	Current stratification, down dip	
		34				N53W	Current stratification, down dip	
		35				N53W	Current stratification, down dip	
		36				N53W	Current stratification, down dip	
		37				N54W	Current stratification, down dip	
		38				N54W	Current stratification, down dip	
		39				N54W	Current stratification, down dip	
		40				N54W	Current stratification, down dip	
	10	1	N47E	90	N43W		Substratal lineation	
		2	21S	90	N43W		Substratal lineation	
		3		90	N43W		Substratal lineation	
		4		90	N43W		Substratal lineation	
		5		90	N43W		Substratal lineation	
		6		90		N42W	Current stratification	
		7		90		N42W	Current stratification	
		8		90		N42W	Current stratification	
		9		90		N43W	Current stratification	
		10		90		N43W	Current stratification	
		11		90		N43W	Current stratification	
		12		90		N43W	Current stratification	
		13		90		N44W	Current stratification	
		14		90		N44W	Current stratification	
		15		90		N44W	Current stratification	
Falli-Hölli Section	15	1	N76E	20SE	N84W		Groove-cast lineation	12
		2	27S	40SE		N64W	Current stratification, fair	
		3		50SE		N54W	Current stratification, down dip	
		4		50SE	N36E		Ripple marks	
	16	1	N76E	5E	N81E		Load-cast lineation	
		2	40S	30E		N74W	Current stratification, from west**	
		3		0	N76E		Convolute bedding	
		4		50SE		N54W	Current stratification	
		5		25SE	N79W		Substratal lineation	
	17	1	N84E	25SE	N71W	N71W	Groove-cast lineation w/ flute casts	
		2	45S	90	N 6W		Load-cast lineation	
		3		70SSE	N26W		Substratal lineation	
	18	1	N57E	70SE	N53W		Substratal lineation	
		2	50S	70SE		N53W	Current stratification	
	19	1	N65E	70SE	N45W		Substratal lineation	

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
Falli-Holli Section	19	2	36S	70SE	N45W		Substratal lineation	12
		3		70SE		N45W	Current stratification**	
		4		70SE	N45W		Parting lineation	
		5		60SE	N55W	N55W	Groove-cast lineation w/ flute casts	
		6		60SW	N 5E		Load-cast lineation	
		7		75SSE	N40W	N40W	Groove-cast lineation w/ flute casts	
	20	1	N51E 62S	90	N39W		Substratal lineation	
	21	1	N57E	67SE	N56W		Parting lineation	
		2	55S	70SE	N53W		Load-cast lineation	
		3		20E	N77E		Groove-cast lineation	
		4		75SE	N48W	N48W	Groove-cast lineation w/ flute casts	
		5		75SE	N48W	N48W	Groove-cast lineation w/ flute casts	
		6		75SE	N48W		Groove-cast lineation	
		7		70SE	N53W		Parting lineation	
	22	1	N60E 62S	10E	N70E		Load-cast lineation	
	23	1	N44E	70S	N66W		Groove-cast lineation	
		2	23S	70S	N66W		Groove-cast lineation	
		3		70S	N66W		Groove-cast lineation	
		4		70S	N66W		Groove-cast lineation	
		5		70S	N66W		Groove-cast lineation	
		6		70S	N66W		Groove-cast lineation	
		7		70S	N66W		Groove-cast lineation	
		8		70S	N66W		Groove-cast lineation	
		9		70S	N66W		Groove-cast lineation	
		10		70S	N66W	N66W	Groove-cast lineation w/ flute casts	
		11		70S	N66W	N66W	Groove-cast lineation w/ flute casts	
		12		70S	N66W	N66W	Groove-cast lineation w/ flute casts	
		13		90		N46W	Current stratification	
	24	1	N67E	25E	N88W		Groove-cast lineation	
		2	20S	40ESE	N73W		Groove-cast lineation	
		3		10WSW	N57E		Groove-cast lineation	
		4		30ESE	N83W		Groove-cast lineation	
	25	1	N79E 57N	82N	N 3W		Groove-cast lineation	
	26	1	N65E	60NE	N 5E		Groove-cast lineation	
		2	52N	60NE	N 5E	N 5E	Groove-cast lineation w/ flute casts	
	27	1	N52E 41S	50SE		N78W	Current stratification	
	28	1	N38E	32S	N 6E		Groove-cast lineation	
		2	24SE	50S	N12W		Groove-cast lineation	
		3		60S	N22W		Groove-cast lineation	
		4		60S		N22W	Current stratification**	
		5		70SE	N72W		Substratal lineation	
		6		70SE	N72W		Substratal lineation	

TABLE 3 (Cont.)

1	2	3	4	5	6	7	8	9
Section	Station	Observation number	Strike and dip of bedding (Ov = Overturned)	Pitch (rake) of current structure, with direction	Bearing of lineation, rotated into horizontal, where sense from which flow came is unknown	Bearing to source of current, rotated into horizontal, where sense from which flow came is known	Type of directional current structure and remarks	Plotted on Figure
Falli-Hölli Section	28	7	N38E 24SE	90		N52W	Current stratification	12
	29	1	N46E	75S	N29W		Substratal lineation	
		2	25SE	75S	N29W		Substratal lineation	
		3		75S	N29W		Substratal lineation	
		4		75S	N29W		Substratal lineation	
		5		75S	N29W		Substratal lineation	
		6		75S		N29W	Current stratification, down dip**	
		7		75S		N29W	Current stratification, down dip**	
		8		75S		N29W	Current stratification, down dip**	
		9		75S		N29W	Current stratification, down dip**	
		10		75S		N29W	Current stratification, down dip**	
	30	1	N88W	0	N88W		Groove-cast lineation	
		2	14S	0	N88W		Groove-cast lineation	
		3		0	N88W		Groove-cast lineation	
		4		0	N88W	N88W	Groove-cast lineation w/ flute casts	
		5		0	N88W	N88W	Groove-cast lineation w/ flute casts	
		6		20W	N72E	N108W	Groove-cast lineation w/ flute casts	

Note: Data in columns 4 and 5 are exactly as measured in the field; some data in columns 6 and 7 however, after being "turned-back" to the horizontal, have been spread out carefully so that they could be plotted conveniently on Figures 8–12.

** Bearing plotted on Figures 8–12 taken from the associated lineation and not from the current stratification.

similar argument, Flysch with a source to the northwest, such as the Gurnigel sandstone, can not be considered as grading laterally into these Helvetic formations. Therefore the Ultrahelvetic Flysch of the External Prealps does not grade laterally into Helvetic rocks of the same age. It must have accumulated in an entirely different region of sedimentation.

In summary, the reconstruction of the depositional setting of the Gurnigel sandstone suggests that the formation accumulated in a relatively sheltered marine trough with a landmass not far to the northwest which contributed clasts of Ultrahelvetic crystalline and sedimentary rocks, an idea independently suggested by Tercier (unpublished geologic profiles). This landmass, which may have appeared as a narrow island or chain of islands, must have been near the northwestern margin of the Ultrahelvetic region, and as it was raised sharply during the Paleocene and Eocene, it shed fragments of older Ultrahelvetic rocks southeastwards into the near-by sea to form the Gurnigel Flysch. Debris washed to the northwest, including clasts of red granite (Habkerngranit), may be represented as the Assilinengrünsand found between Seewen and Schwyz in shallow-water beds (Heim, 1908). Perhaps the swell was formed during the Lias, since clasts of the same composition as in the Flysch occur locally in the Ultrahelvetic Lias, Dogger, and Malm, but it did not again protrude above the sea until the Eocene. The Cretaceous rocks, and to some extent the Jurassic on the northeast, show significant facies gradations between the Helvetic and Ultrahelvetic regions (Heim, 1922), which have allowed much of our understanding of the tectonics to be worked out. During the Eocene and Oligocene the ridge may have been prominent for a while but afterwards it was probably dwarfed by swells rising to the southeast which shed aprons of Flysch debris into the sea as the

orogeny came to a culmination. Today the roots possibly lie buried beneath the Pennine nappes. The Ultrahelvetic sediments southeast of the swell may have been stripped from their basement during Oligocene and Pliocene deformations, and perhaps were squeezed or glided from their resting place, so that they now form the Ultrahelvetic nappes far to the northwest.

Conclusions

This study of the Prealpine Flysch confirms that oriented current structures can tell much about paleogeography and the area of sedimentation. In rocks laid down by turbidity currents, the structures yield data on the slope direction of the sea floor, and, in turn, on the direction from which much of the sediment has come. It is desirable to study such structures from many rock sequences in more detail, so that we can use them to their maximum advantage. In the search for oil and water, knowledge from current structures, in combination with other data, may help find areas of greater sand thickness and permeability. They also may help subdivide stratigraphic sections for mapping.

The compilation of oriented observations from the External Prealps shows that the Gurnigel Flysch was laid down in a fairly deep marine trough with a landmass not far to the northwest. This conclusion requires a Paleocene-Eocene swell in the northwestern part of the Ultrahelvetic sedimentation region, and bordering the Helvetic region. With the culminations of the Alpine orogeny in the Oligocene and Pliocene, the Gurnigel Flysch moved far to the northwest within the Ultrahelvetic nappe-complex.

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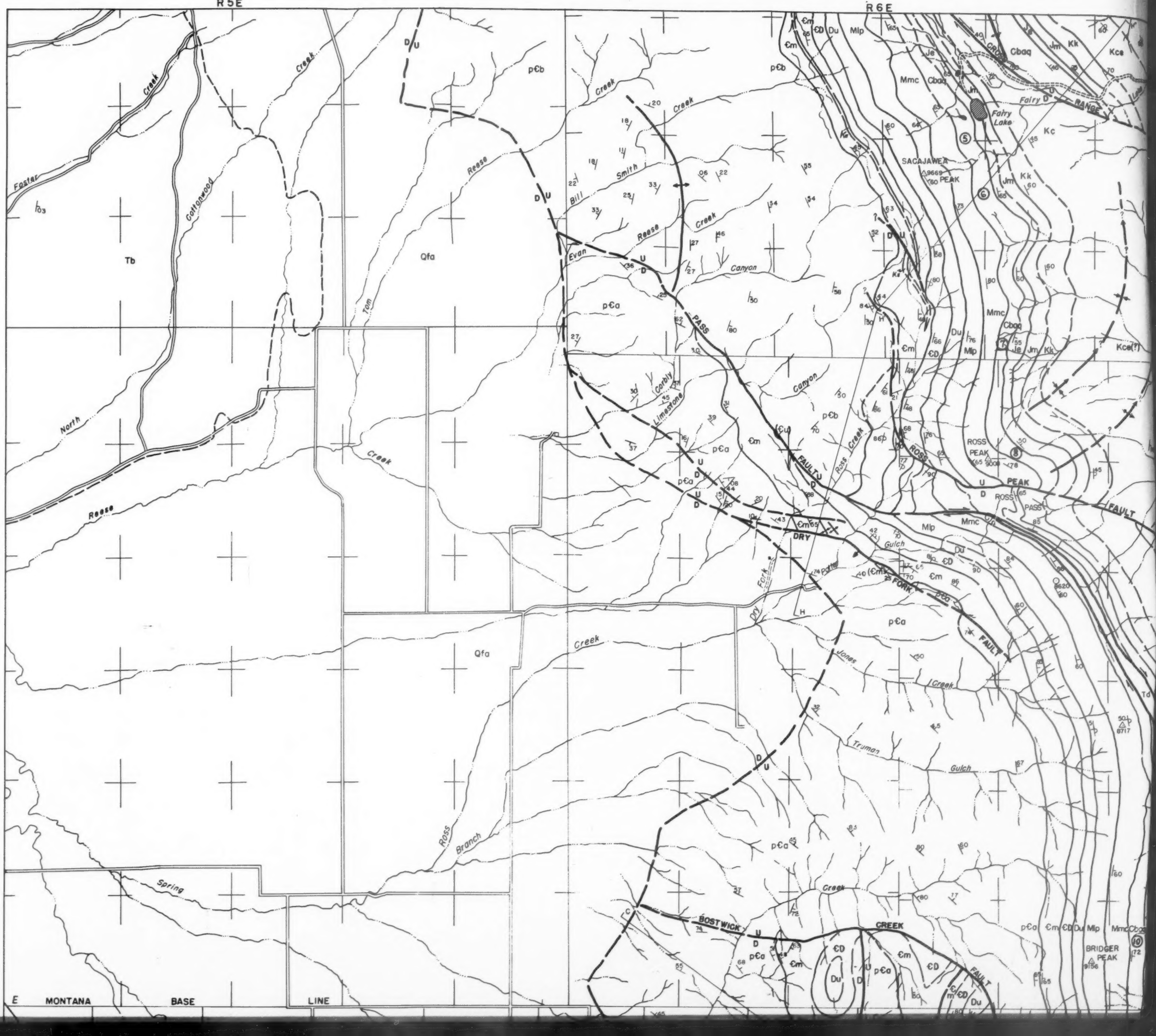
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Carbly Limestone Creek
Ross Fork Creek
Dry Fork Creek
Jones Creek
Truman Gulch
Ross Branch
Fairy Lake
SACAJAWEA PEAK
ROSS PEAK
BRIDGER PEAK
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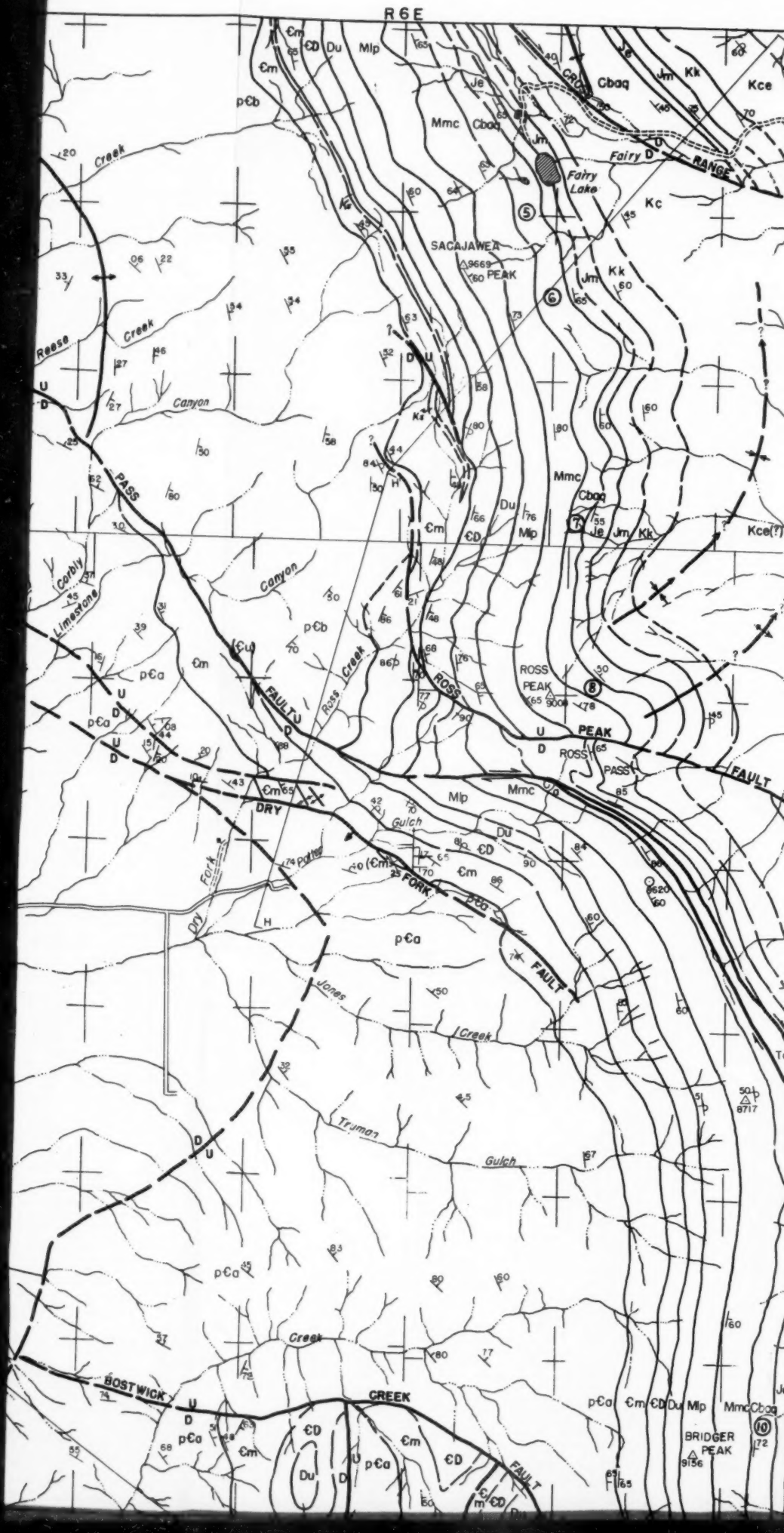
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Dry
Fork
25 FORK
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Jones
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Truman
Gulch
Creek
BOSTWICK
CREEK
FAULT
BRIDGER
PEAK
9156
Fairy
Lake
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RANGE
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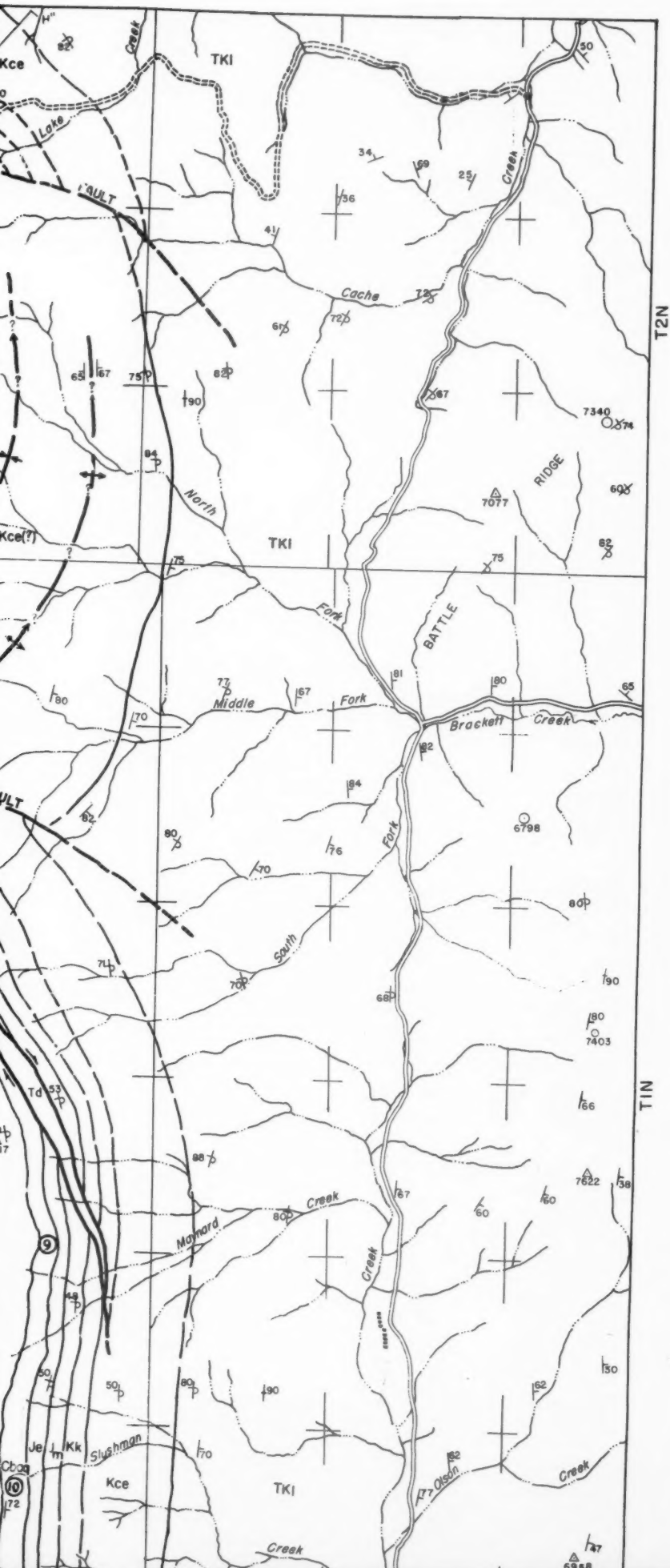
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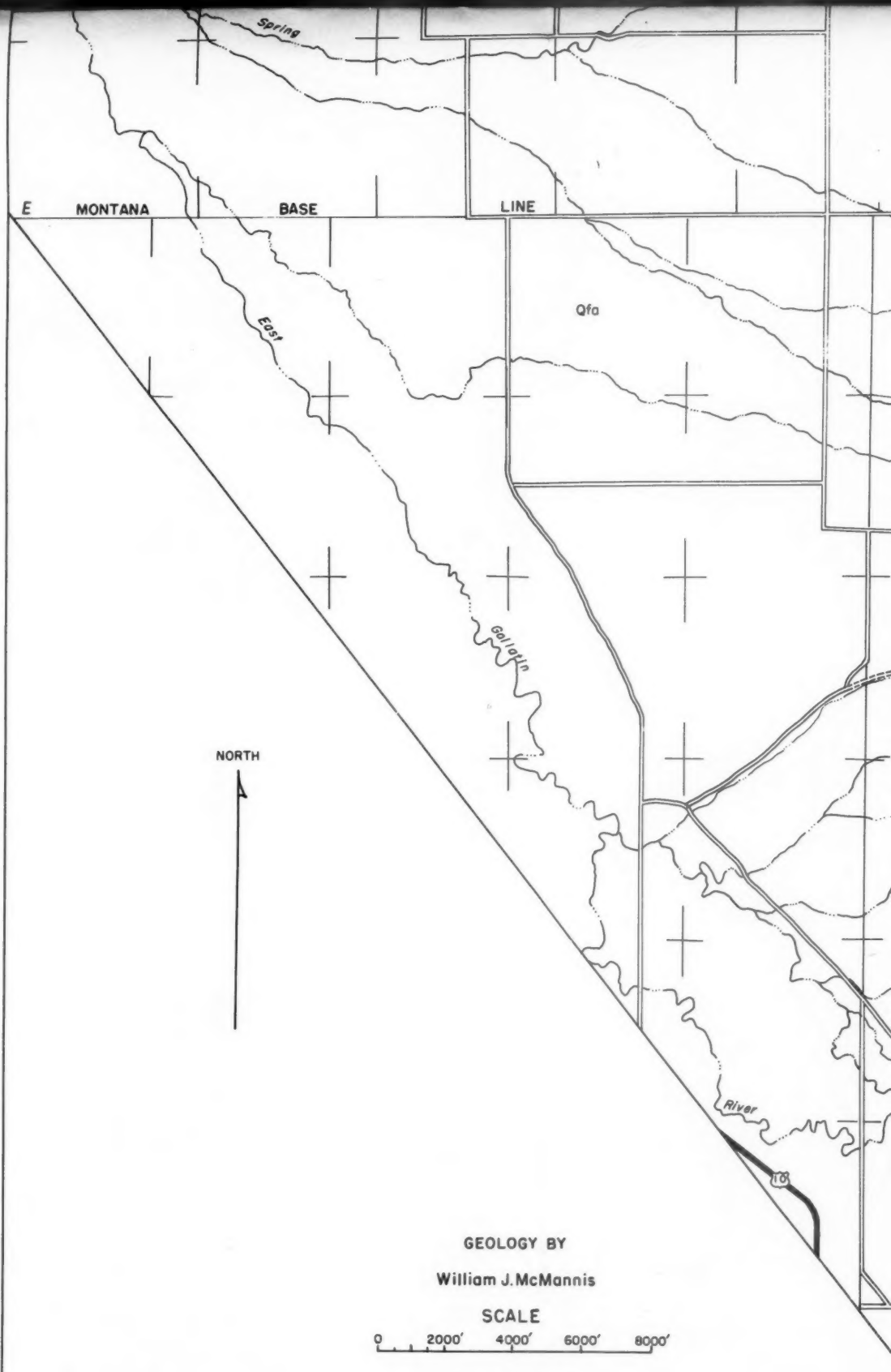


GEOLOGY BY

William J. McMannis

SCALE

0 2000' 4000' 6000' 8000'

LEGEND

SEDIMENTS

QUATERNARY	Qfa	Fan and Alluvial Gravels
TERTIARY	Tb	Basin Deposits Oligocene-Miocene
UPPER CRETACEOUS	TKl	Livingston Formation
	Kce	Colorado and Eagle Formation
LOWER	Kk	Kootenai Formation
JURASSIC	Jm	Morrison Formation
	Jmc	
	Je	Ellis Group (includes chert breccia at base)
MISSISSIPPIAN-PENNSYLVANIAN	Cbaq	Quadrant Formation Amsden Formation Big Snowy Group
MISSISSIPPIAN	Mmc	Mission Canyon Formation
	Mlp	Lodgepole Formation
DEVONIAN	Du	Sappington Formation (Devonian?) Three Forks Formation Jefferson Formation

SYMBOLS

FORMATIONAL CON

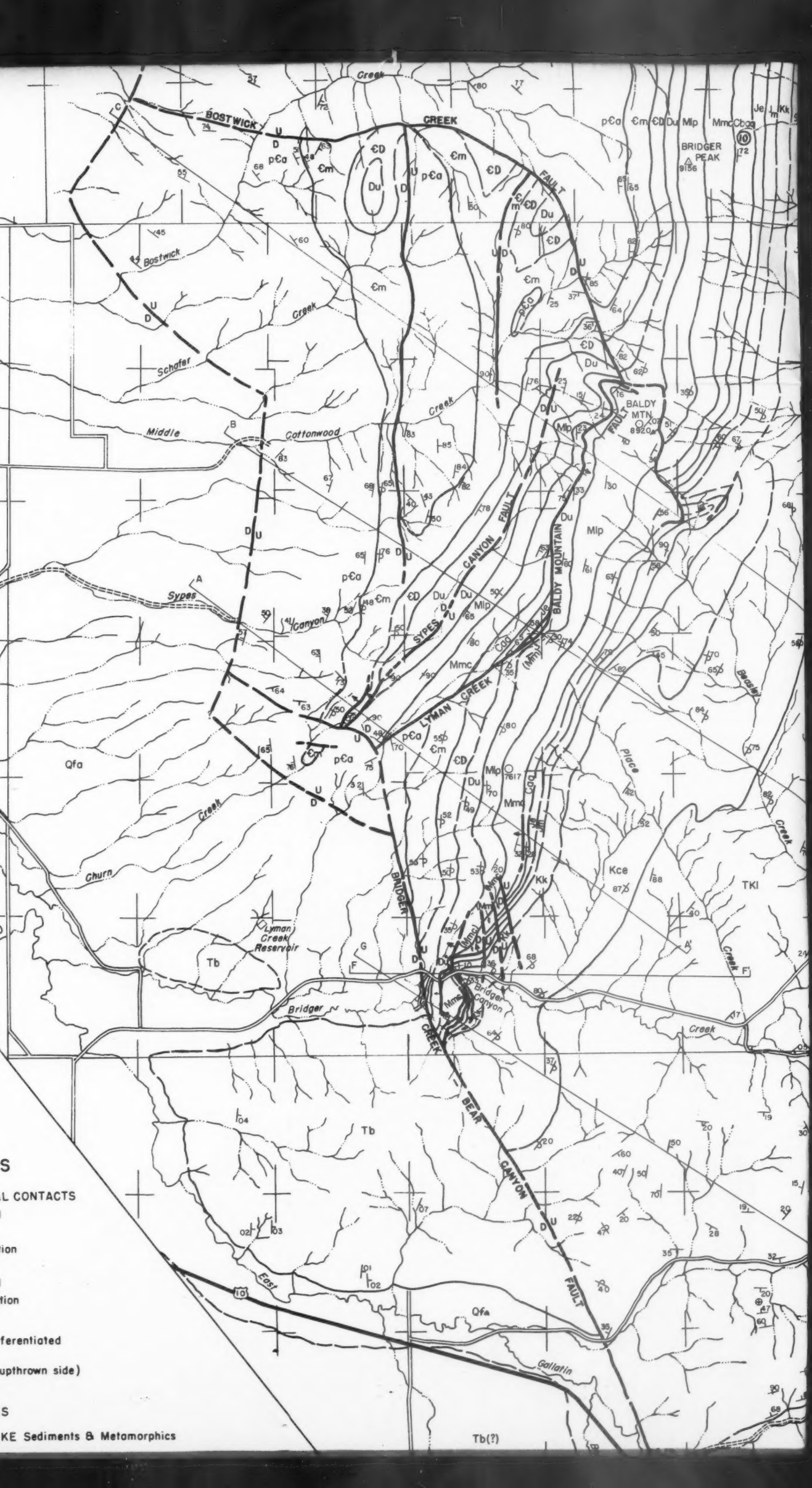
- Definitely Located
- Fairly Certain
- Approximate Position

FAULTS

- Definitely Located
- Approximate Position
- Inferred
- U / D — High Angle, undifferenti
- Thrust (arrow on upthro
- Strike-slip

FOLD AXES

DIP & STRIKE Se



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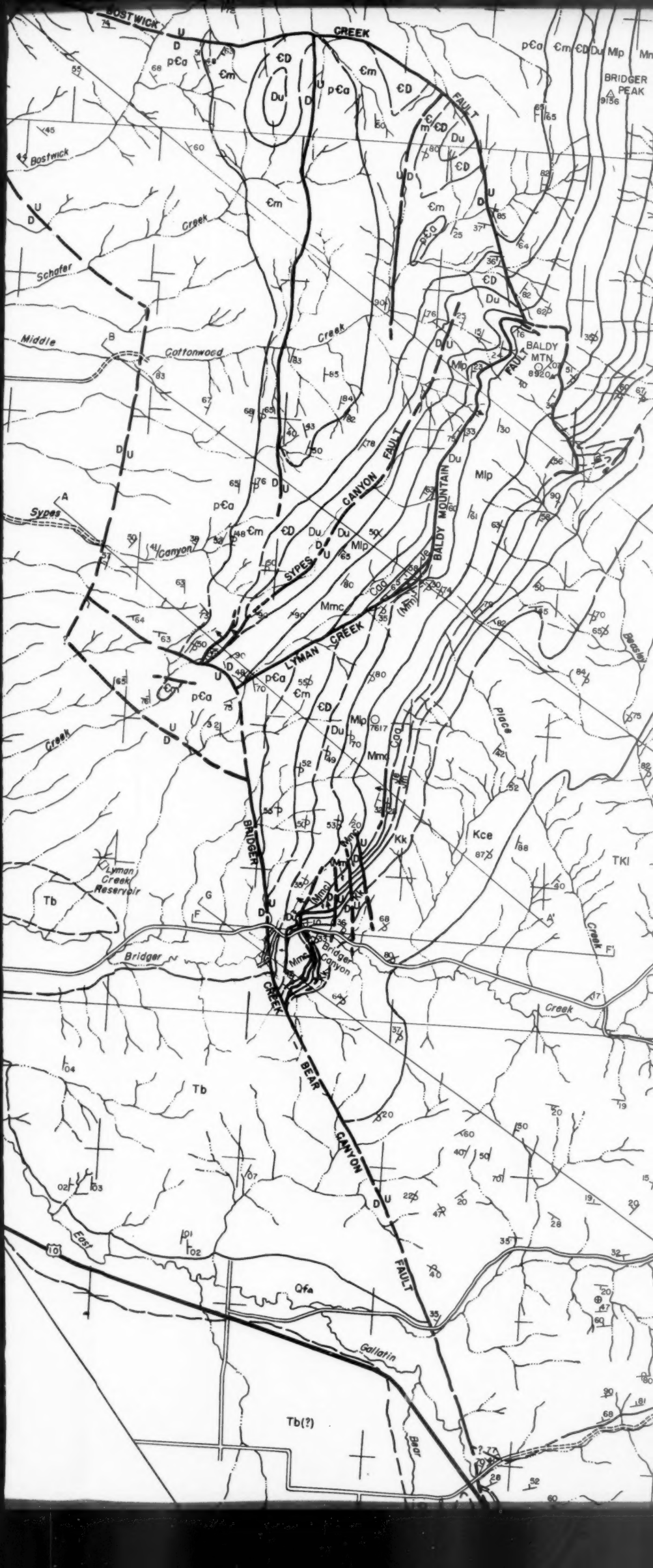
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BRIDGER PEAK
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FAULT
Bostwick
Creek
Schafer
Middle
Cottonwood
Creek
BALDY MTN.
8920
FAULT
CANYON FAULT
BALDY MOUNTAIN
Sypes
Canyon
SYPES
LYMAN CREEK
Beasley
Place
Creek
BRIDGER
Lyman Creek Reservoir
Bridger
Canyon
CREEK
Creek
BEAR CANYON FAULT
East
Gallatin
Bear
Tb
Tb(?)
Qfa
Kce
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Mmc
Mlp
Du
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A
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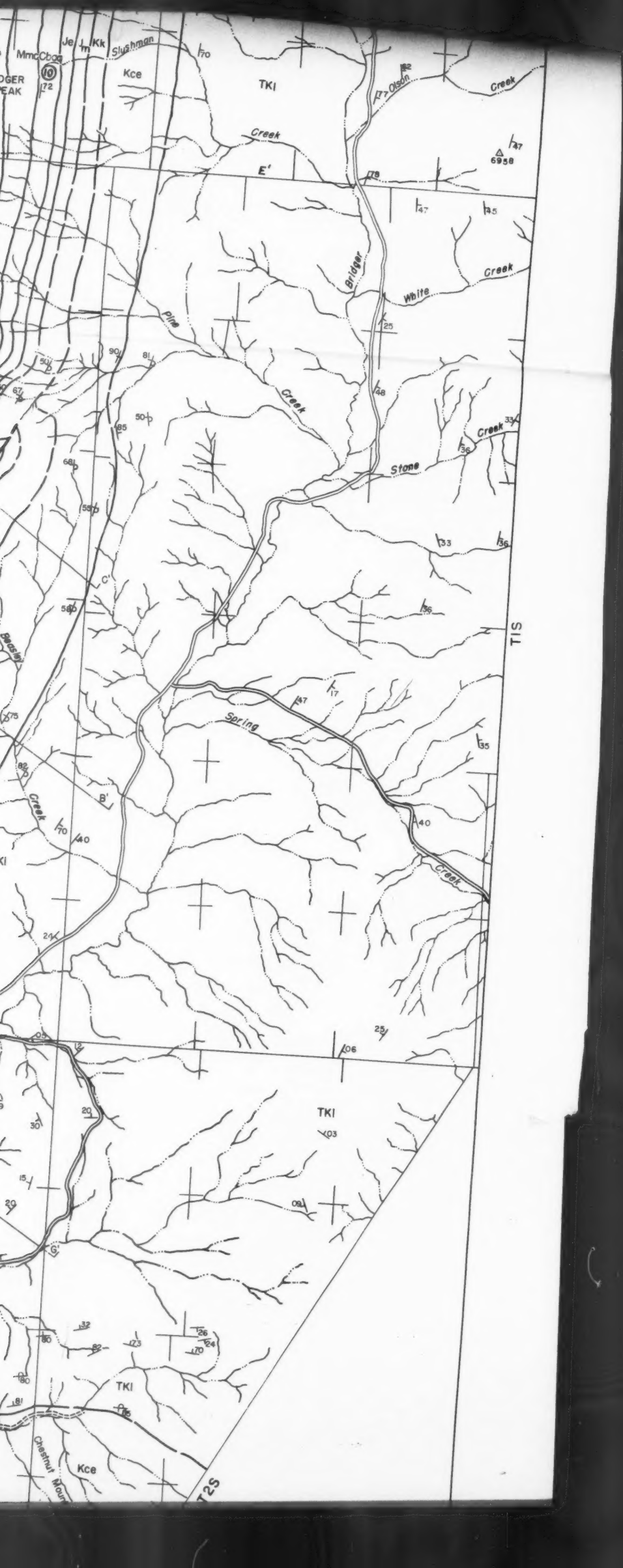
G


Kce
TKI
Creek
Olson
Bridger
White
Pine
Stone
Spring
Chestnut
E'
C'
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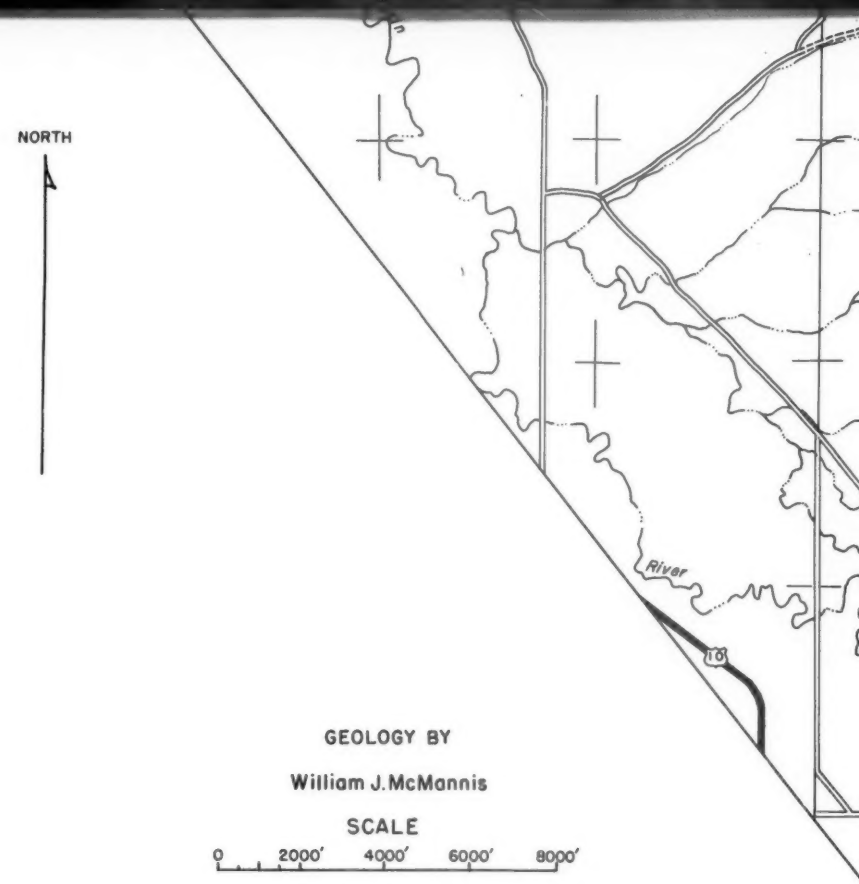
G'
T1S
T2S
6958



GEOLOGY BY

William J. McMannis

SCALE

0 2000' 4000' 6000' 8000'

LEGEND

SEDIMENTS

QUATERNARY	Qfa	Fan and Alluvial Gravels
TERTIARY	Tb	Basin Deposits Oligocene-Miocene
UPPER CRETACEOUS	TKl	Livingston Formation
	Kce	Colorado and Eagle Formation
LOWER CRETACEOUS	Kk	Kootenai Formation
JURASSIC	Jm	Morrison Formation
	Jmc	
	Je	Ellis Group (Includes chert breccia at base)
MISSISSIPPIAN-PENNSYLVANIAN	Cbaq	Quadrant Formation Amsden Formation Big Snowy Group
MISSISSIPPIAN	Mmc	Mission Canyon Formation
	Mlp	Lodgepole Formation
DEVONIAN	Du	Sappington Formation (Devonian?) Three Forks Formation Jefferson Formation
UPPER CAMBRIAN	€D	Maywood (?) Formation Snowy Range Formation Maurice Formation
MIDDLE CAMBRIAN	€m	Park Formation Meagher Formation Wolsey Formation Flathead Formation
PRE-CAMBRIAN	p€b	Belt Arkoses
	p€a	Archean (?) Metamorphics
PALEOCENE UPPER CRETACEOUS	Td	Composite Dike
	Ks	Composite Sill

SYMBOLS

FORMATIONAL CON

- Definitely Located
- Fairly Certain
- Approximate Position

FAULTS

- Definitely Located
- Approximate Position
- Inferred
- U / D — High Angle, undifferenti
- Thrust (arrow on upthrow
- Strike-slip

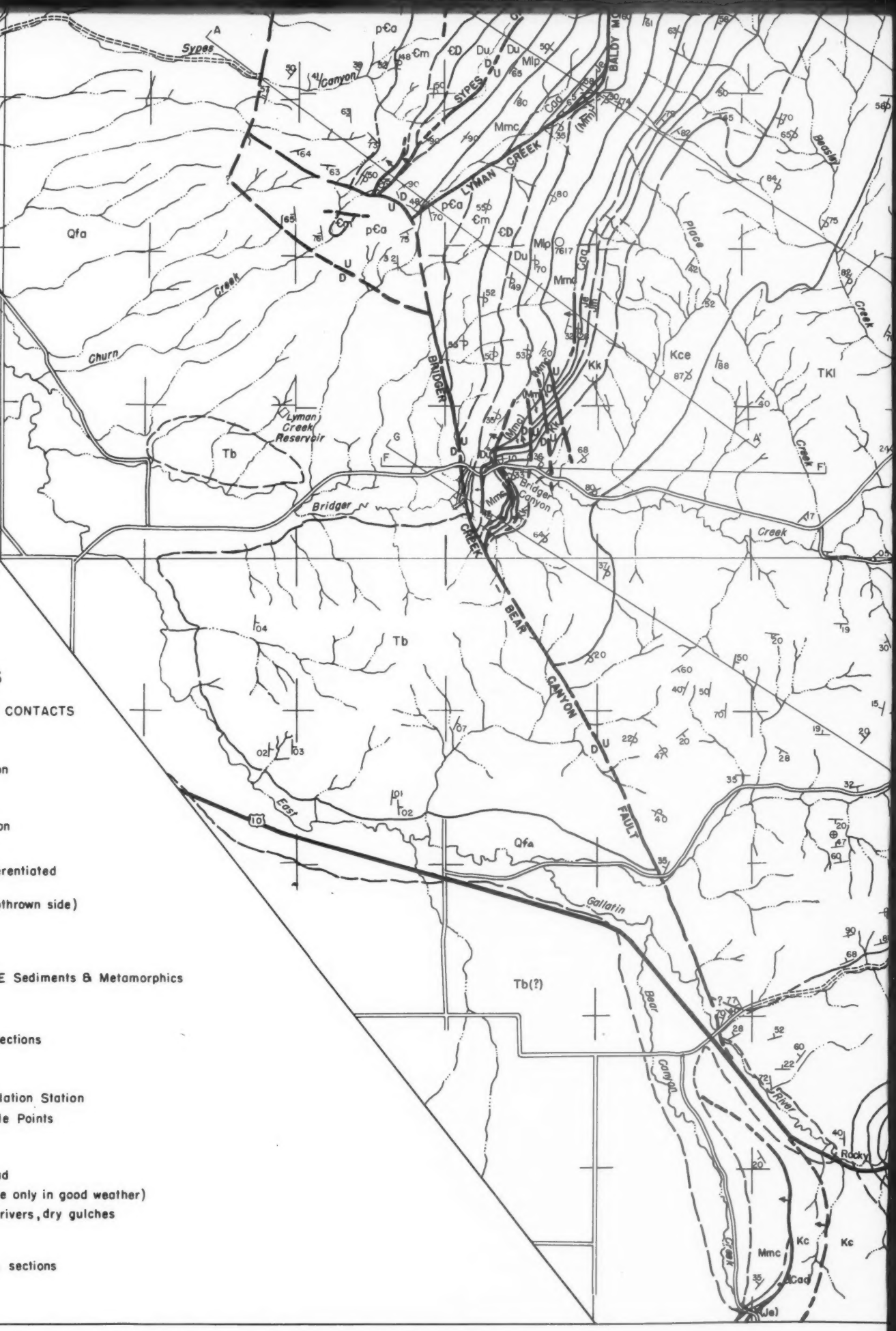
FOLD AXES

DIP & STRIKE Se

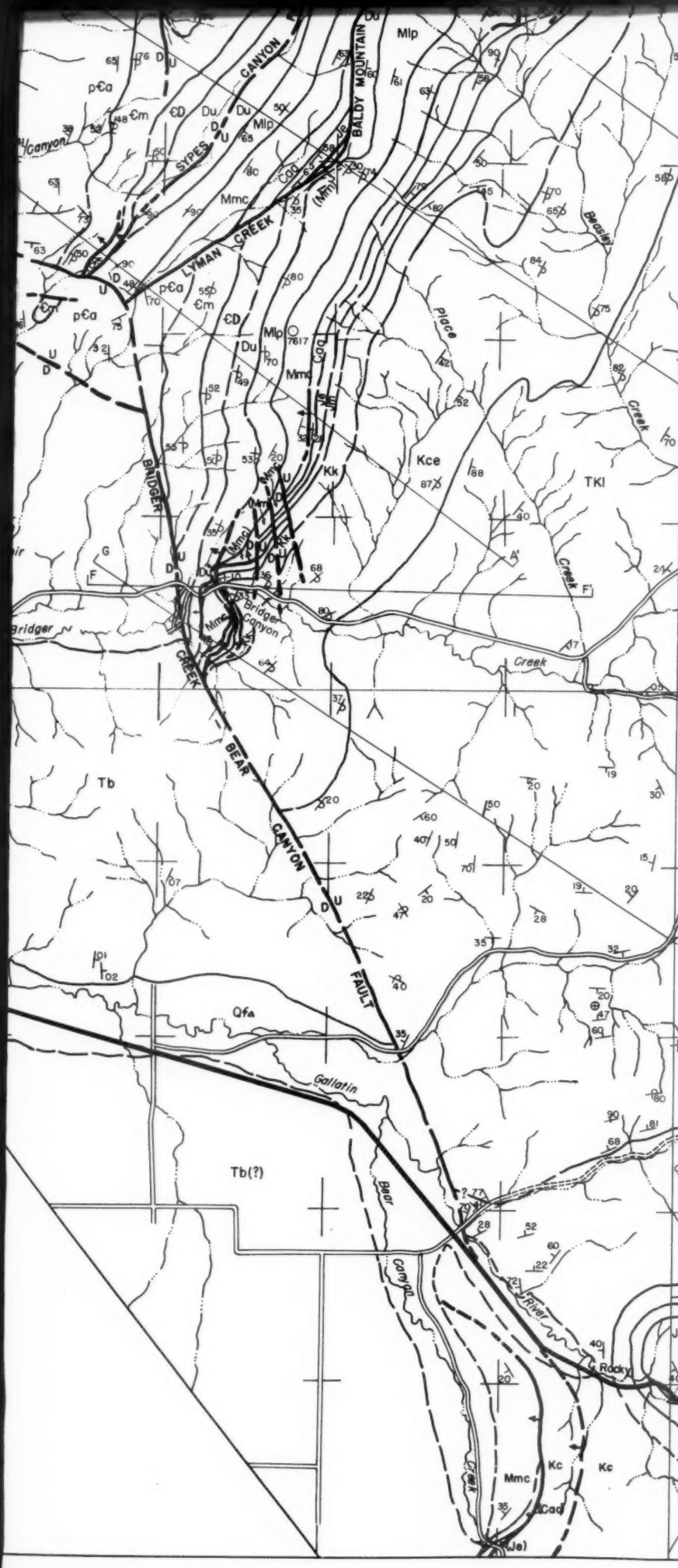
- 22 — Upright beds
- 75 — Overturned beds
- A A' — Location of Cross Sectio
- Section Corner
- △9125 — Secondary Triangulation
- ⊙8756 — Elevation of Notable Po

ROADS

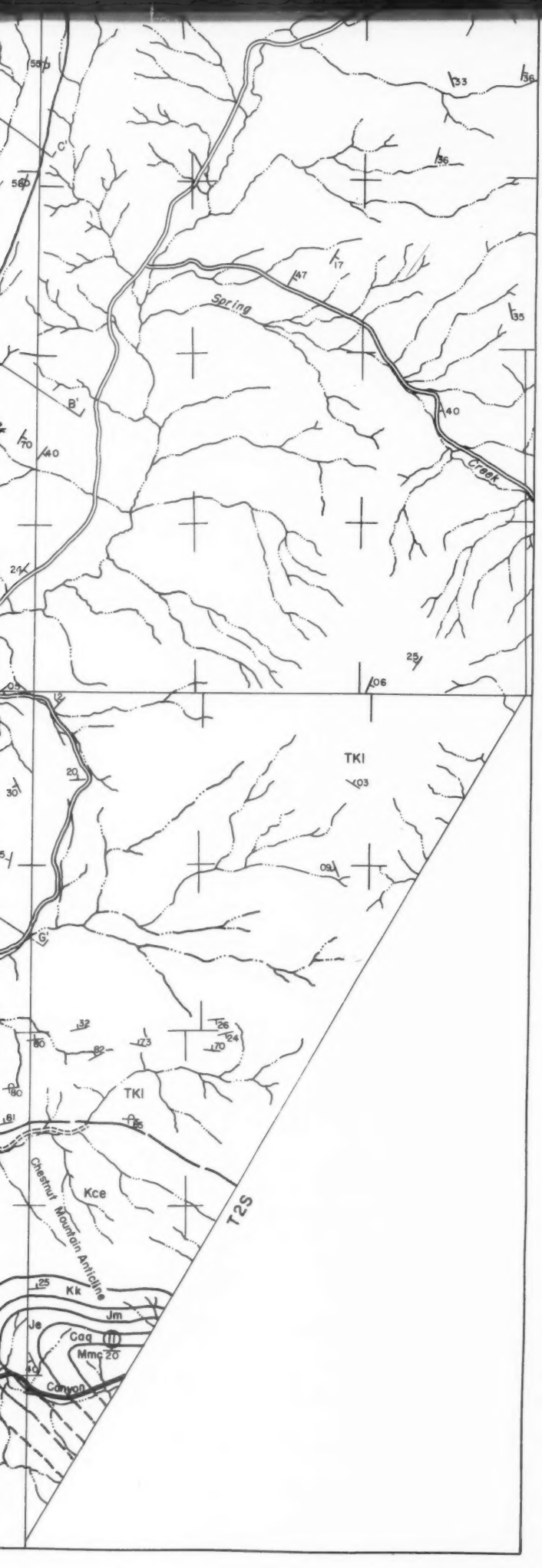
- Blacktop Highway
- Gravel or Dirt Road
- Secondary (passable onl
- Drainage: streams, river
- Lakes
- (9) — Big Snowy-Amsden sect



GEOLOGIC MAP OF BRIDGER RANGE, MONTANA (South Part)



RANGE, MONTANA (South Part)


T1S
T2S
Spring
Creek
TKl
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Chestnut Mountain Anticline
Kk
Jm
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Canyon
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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 1385–1430, 5 FIGS., 8 PLS. NOVEMBER 1955

GEOLOGY OF THE BRIDGER RANGE, MONTANA

BY WILLIAM J. MCMANNIS

ABSTRACT

In and near the Bridger Range area north and east of Bozeman, Montana, in Gallatin County, about 27,000 feet of sedimentary rocks, from Beltian to Recent, is exposed. Strata of known Ordovician, Silurian, Permian, and Triassic age are not present. Middle Cambrian Flathead sandstone is underlain by coarse Beltian arkoses in the northern part of the area and by Archean (?) metamorphics in the southern part. The depositional edge of the Belt arkoses has been fault-controlled. The Sappington formation (Berry, 1943) is separated from the underlying Three Forks formation by a minor disconformity. The Big Snowy group is recognized in the Bridger Range for the first time. The problem of the Livingston formation is reviewed, and it is apparent that the andesitic phases of these Montanan, Lancian, and Paleocene sediments can be used to interpret the tectonic history of the Crazy Mountain basin and adjacent mountain areas.

Analysis of the structural features of the Bridger Range and adjacent areas indicates two phases of folding and faulting. One developed generally north-trending structures in mid-Paleocene time, and the other developed north-northwest trending structures probably in latest Paleocene or early Eocene.

In Oligocene time isostatic arching produced normal faults on the west side of the Bridger Range. The range was lifted relative to the Gallatin Valley, and the east-flowing drainage was impeded so that sediments accumulated in the basins. These movements and deposition have continued sporadically to the present.

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INTRODUCTION

The Bridger Range area of southwestern Montana provides critical evidence with respect to local and regional Northern Rocky Mountain tectonic history. The area is on the boundary between complex structures of the mountains proper and relatively simple structures of the central Montana region. This area was studied as part of the regional mapping program conducted by members of Princeton University in conjunction with the Yellowstone-Bighorn Research Association. The writer aimed to construct an accurate geologic map, collect stratigraphic and structural data, and study critical physiographic relations. These objectives were partially attained during the summers of 1950–1951, and additional work was done in July and August 1952. Further detailed work remains to be done in this and near-by strategic areas.

The Bridger Range lies directly north and slightly east of Bozeman, Montana (Fig. 1), in Ts. 1 and 2 S., and 1–3 N., and in Rs. 5, 6, and 7 E. The range extends from Bridger Canyon northward in a gently curving arc for 23 miles to Blacktail Mountain. It forms part of the southwestern rim of the Crazy Mountain Basin and is bounded on the west by the Gallatin Valley, which has been down-dropped by Tertiary faulting. The range is terminated on the north by the Maudlow Basin and is cut off on the south by a major oblique fault.

The mapping covered the Bridger Range and Chestnut Mountain areas and adjacent portions of the Gallatin Valley and Crazy Mountain Basin (Pls. 1 and 2). Plate 1 shows the geology of the area mapped by the writer, except for that in the immediate vicinity of Chestnut Mountain which is slightly altered from that of Skeels (1939). Plate 2 is a composite of reconnaissance and photo geology by the writer and of mapping by Loufek, Jeanes, and Klemme, of Princeton University.

In 1893 Weed studied the Laramie (Eagle) and Livingston formations along the west rim of the Crazy Mountain Basin. The Bridger Range area was roughly mapped during the preparation of the Livingston Folio (Iddings and Weed, 1894) and the Three Forks Folio (Peale, 1896). Stone and Calvert (1910) examined the Eagle coal-bearing beds and the Livingston formation along the east flank of the Bridger Range; their interpretations of age and stratigraphic relations differed from Weed's. Condit (1919) sampled bituminous shales from what are here called the Sappington formation and the Big Snowy group.

Skeels (1939) mapped the southern extremity of the Bridger Range area. Dorf and Lochman (1940) briefly examined the Cambrian strata in the southern part of the range.

Cobban (1945) visited the southern part of the area and redesignated the type section for the Ellis formation. Gardner *et al.* (1946) published measured sections of upper Paleozoic and Mesozoic rocks from three localities in the map area—Rocky Canyon, Bridger Canyon, and Fairy Lake.

Klemme (Unpublished Thesis, Princeton Univ., 1949) mapped southward from his Sixteen Mile Canyon area as far as Lat. 46° N. in the northern portion of the Bridger Range (Pl. 7). Pardee (1950) devotes a few short paragraphs to late Cenozoic faulting along the west side of the range.

During the summers of 1950 and 1951 the geology was plotted on aerial photographs with

a scale of 1/24,000; 20-foot contour topographic sheets made by the Bureau of Reclamation were available for the entire map area. Geological data were transferred from aerial photographs to the topographic sheets with the Fairchild Recto-planograph and by eye through comparison of drainage and topographic features. Base maps and geologic maps were compiled from individual topographic sheets.

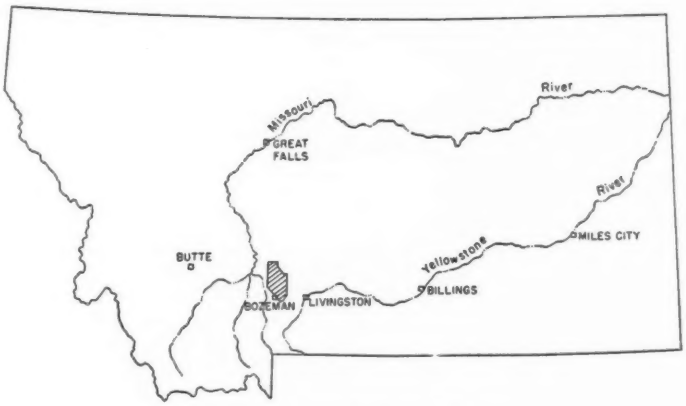


FIGURE 1.—INDEX MAP OF MONTANA

ACKNOWLEDGMENTS

The writer wishes to acknowledge the assistance of many individuals and organizations including Princeton University and The Geological Society of America for financial assistance; A. R. Palmer, J. B. Smedley, G. A. Cooper, A. L. Bowsher, L. A. Thomas, G. L. Jepsen, B. Howell, and E. Dorf, for paleontological examination and identification; O. M. Hackett and F. A. Swenson, for furnishing data on the depth of Recent and Tertiary fill in the Gallatin Valley; M. E. Loufek, H. Jeanes, F. W. Rossetter, G. G. Anderman, and A. Pinsak, for assistance in field mapping; W. R. Lowell for helpful criticism and discussion in the field concerning mapping problems; W. T. Thom, Jr., and other members of the Geology Department of Princeton University for discussion and critical reading of part or all of the original manuscript.

STRATIGRAPHY

General

The rock units (Table 1) range from Archean (?) to Recent and are about 27,000 feet thick, not including the thickness of the metamorphic series. All the Paleozoic systems are represented except the Ordovician, Silurian, and probably the Permian. Triassic strata are absent in the Bridger Range.

All formations from the Archean (?) through the Ellis group are well exposed; younger strata are not. Lithology and age of the conglomerates in the upper part of the Livingston formation are stressed because of their immediate bearing upon the phases of Laramide deformation in this area.

Precambrian

Archean (?).—In the southern half of the Bridger Range extensively exposed Precambrian metamorphics are dominantly black, white, and some pink, quartz-feldspar gneisses. Garnet-amphibole schists are common, with gradations from nearly pure garnet amphibolites to a garnet-quartz-feldspar gneiss with minor amphibole. Muscovite-and sericite-rich quartzites are found at several localities, as are nearly pure amphibolites. Several beds of dolomitic marble are prominent in the vicinity of Ross Creek and Potter Gulch. Small pegmatites composed of quartz and pink or white ortho-

TABLE 1.—GENERALIZED STRATIGRAPHIC SECTION

Age		Stratigraphic Unit		General Lithology	Thickness Feet
QUATERNARY				Valley fill, alluvium, gravel fans, outwash and morainal material	0–300 +
		unconf.			
TERTIARY		Undifferentiated		Ash, silt, sand, and fanglomerate	0– ?
		unconf.			
				Coarse conglomerate, some andesitic sand lenses Andesitic sandstone, sporadic conglomerate beds	
		—unconf. ?—			
				Coarse conglomerate, some andesitic sandstone	
	?	Livingston fm.		Siltstone, shale, some andesitic sandstones and fresh-water limestones Andesitic sandstone, andesite conglomerates	up to 14,500
CRETACEOUS	UPPER	Eagle sandstone		Salt and pepper sandstone, partly marine	100–600
		Colorado fm.		Black shale, rusty gray-green sandstone and siltstone, minor gray, salt and pepper sandstone Marine	1200–2400
	?				
	LOWER	Kootenai fm.		Basal conglomeratic sandstone and upper sandstone medial red-purple claystone and shale	386–447
		disconf.			
JURASSIC		Morrison fm.		Variegated shale and mudstone, with interbedded rusty calcareous sandstones	110–444
		ELLIS GROUP	Swift sandstone	Yellow calcareous sandstone, basal conglomerate or pebbly zone	50–100
			disconf.		
			Rierdon fm.	Massive grey, oölitic limestone, with overlying shaly beds	0–114
			Sawtooth fm.	Fine-grained dark-gray limestone, interbedded shale. Fragmental limestone with chert pebbles in lower part. Locally at top a red-yellow siltstone	20–145
			disconf.		
JURASSIC (?)				At many places a chert breccia and conglomerate occurs at this horizon	0–26
		disconf.			
PENNSYLVANIAN		Quadrant fm.		Pale-yellow to white, pure quartz sandstone or quartzite, locally calcareous. A few thin, light-gray dolomite beds	40–165
		Amsden fm.		Upper light-gray dolomite with some thin quartz sandstone beds	113–185
				Lower red siltstone unit with some variegated dolomite and impure fossiliferous limestone beds	11–189
	?				
		disconf.			
MISSISSIPPIAN	UPPER	BIG SNOWY GROUP		Cherty limestone Black shaly limestone and black shale	0–171
			local disconf.		
				Locally red silty beds at top Red and yellow sandstone or calcarenite	0–163
				Red siltstone, with a few purple and pale-yellow splotched dolomites Basal dolomite and dolomite or limestone breccia	0–100

TABLE 1.—*Continued*

Age		Stratigraphic Unit		General Lithology	Thickness Feet
MISSISSIPPIAN	LOWER	MADISON GROUP	—disconf.— Mission Canyon fm.	Solution channels, caves at top Massive, light-gray limestone, solution breccias at several horizons	430–950
			Lodgepole fm.	Thin-bedded, yellow to red-stained, fossiliferous limestone Lower dark-gray, thin-bedded, less fossiliferous limestone	750–810
—?—					
DEVONIAN (?)			Sappington fm.	2–3 feet black silty shale Yellow sandstone grading downward into silty and sandy limestone Basal black shale	46–99
—?—			—disconf.—		
DEVONIAN			Three Forks fm.	Gray limestone at top. Yellow siltstone and green shale. Middle ledge-forming, brecciated limestone. Basal evaporite solution breccias, and red and orange nodular limonitic shale	155–156
			Jefferson fm.	Light- and dark-brown, thick-bedded dolomite, dolomitic limestone, and limestone. A few intercalated yellow and pale pink dolomitic siltstone beds.	497–620
			Maywood (?) fm.	Yellow mudstone and siltstone, thin dolomite beds. Grades upward into Jefferson. Red, blocky siltstone, with red-stained brecciated limestone beds in lower part. Basal red, fissile shale.	39–92
—?—			— disconf. —		
CAMBRIAN	UPPER	SNOWY RANGE FORMATION	Sage pebble conglomerate member	Fine-grained, thin-bedded, dense limestone and limestone pebble conglomerate, with interbedded green shale. Basal columnar limestone.	121–204
			Dry Creek shale member	Gray-green fissile shale, with interbedded yellow calcareous siltstone and sandstone	42–76
			Maurice fm.	Massive light- and dark-gray mottled, oölitic limestone. Local reefoid development at base Thin to thick-bedded, gray, edgewise and flat-pebble limestone conglomerate with interbedded green shale Basal ledge-forming massive, oölitic, mottled limestone	363–433
	MIDDLE		Park fm.	Green and maroon fissile shale, with thin-bedded limestone unit at top and, locally, intercalated conglomeratic, arkosic limestone and arkose beds in lower portion	188–192
			Meagher fm.	Thin-bedded, dark-gray, dense limestone with interbedded green shale Middle massive dark-gray dense limestone Thin-bedded, dark-gray, dense limestone with interbedded green and yellow silty shale	368–370
			Wolsey fm.	Green and maroon, micaceous shale with interbedded micaceous sandstone and siltstone. Locally contains conglomeratic arkosic limestone and arkose	152–210
			Flathead fm.	Red, pale orange, and white sandstone, locally quartzitic. Locally contains much feldspar, becoming arkosic. Conglomeratic in places	119–142

TABLE 1.—*Concluded*

Age		Stratigraphic Unit	General Lithology	Thickness Feet
		—unconf.—		
PRECAMBRIAN	ALGONKIAN	Belt series	Coarse, massive, poorly bedded arkoses and conglomeratic arkoses. Very coarse gneiss boulder conglomerate in southwest part of area. Interbedded dark-gray argillite and a few siliceous limestone beds in northern part of area	6000 +
		—Fault—		
	ARCHEAN (?)	Metamorphics	Gneiss, schist, metaquartzite, marble, injection gneiss, numerous pegmatite dikes, and veins.	?
			Approximate total thickness of section	27,000 +

clase, rarely containing small biotite books, occur in the metamorphic terrane. Locally injection gneisses are common; the invading material is granitic, and the host consists of various kinds of amphibole-bearing rocks.

The metamorphics are well banded, and foliation is easily measured. Flowage folds and other minor flexures are common. At normal stratigraphic contacts the attitude of metamorphic foliation and of bedding in the overlying Flathead sandstone is similar. West and southwest of Bridger Peak, the foliation curves around the noses of two plunging folds, nearly paralleling the overlying Cambrian strata. This suggests that the basement rocks have also been folded and faulted during more recent orogenic activity. The rock types in the metamorphic terrane suggest that most of this series was sedimentary, particularly the marbles, muscovite-rich quartzites, and garnet-rich amphibolites and gneisses.

In southwestern Montana two series of metamorphics, the Pony below and the Cherry Creek above, have been named and described (Tansley *et al.*, 1933, p. 8–12). Workers disagree on the validity of such a subdivision. Little is known of the relationship and ages of the metamorphics in southwestern Montana. For these reasons the term Archean (?) is used here.

Belt Series.—In the northern half of the Bridger Range more than 6000 feet of coarse clastics forms part of the normal Belt Series present north of this area. The rocks vary from extremely coarse conglomerates, composed of cobbles, boulders, and blocks of gneiss, quartzite, and schist, to coarse arkosic sandstones. The conglomerate matrix is about the same composition as the sandstones.

The sandstones contain much quartz, microcline, and orthoclase, with some plagioclase, plus a binding material of dark-green chlorite and altered biotite and amphibole. Invaginate outlines of quartz and feldspars, a feature inconsistent with sedimentary rounding, suggest slight alteration or metamorphism. (*See also* Merrill, 1893, p. 47–54.)

The rock weathers dark brownish green, and in places reddish. Fresh fractures are always moderately dark gray green. Much of the bedding is indistinct because the clastics are massive. Dips are difficult to ascertain because of this and because individual grains or particles are unoriented. Ripple marks, cross-bedding, mud cracks, and rain prints are also lacking, suggesting rapid deposition. Seemingly the water into which these clastics poured was much deeper than that of the Belt basin north of this area, and/or subsidence was rapid.

The coarse conglomerates are confined to the southwesternmost exposures; however, some gneiss pebbles and more quartz pebbles are common throughout the area. In the north beds of dark-gray argillite are interbedded with the arkoses, as are a few thin beds of siliceous limestone. In general coarse clastics diminish both to the north and to the east. This indicates a source to the southwest.

No contacts between the arkoses and the older metamorphic rocks were seen. The only visible contact is along the trace of Pass fault (Pl. 1). This fault marks the southern extent of

Beltian arkoses and the northern limit of Archean (?) metamorphics in the Bridger Range. This fault contact must be near the original depositional edge of the Beltian conglomerates.

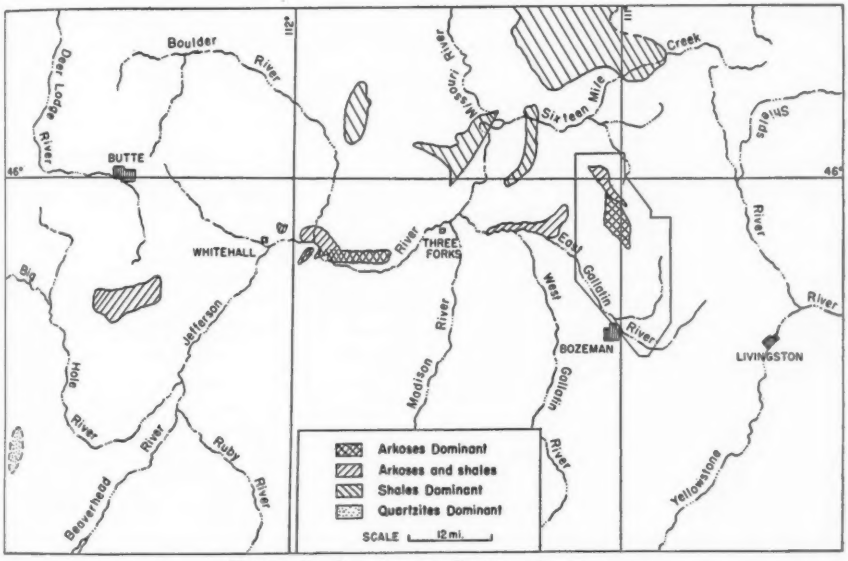


FIGURE 2.—LOCATION OF BELTIAN ARKOSES

Pass fault can be traced from Bill Smith Creek to Ross Pass and thence south-southeast across the headwaters of Maynard Creek on the east side of the Bridger Range (Pl. 1). Apparent displacement near the southeast end of the fault, along the dike, is less than 500 feet but increases gradually from Ross Pass westward to Dry Fork of Ross Creek where it is at least 1500 feet. Thence northwestward the displacement increases to about 5000 feet where the Flathead sandstone is in contact with the lower part of the thick Belt arkose sequence.

Farther northwest along the fault, the apparent stratigraphic displacement is opposite that described. The Laramide movement on this fault must have been opposite that along the same fault in Precambrian time (Fig. 5). Faulting must have controlled deposition of the coarse Beltian conglomerates. Nearly all workers have assigned the conglomerates a Beltian age. Recent studies have further clarified the relationships of these beds. Klemme (Unpub. Thesis, Princeton Univ., 1949) noted coarse arkoses interbedded with argillites or shales of type Belt appearance in the Horseshoe Hills area near Logan. He suggests that the coarse facies "interfingers laterally and grades vertically" with the Spokane shale of the Belt Series. The base of the Belt Series is not exposed; deeper underlying portions, resting on the metamorphics, may be a shoreward facies of the next older formation, the Greyson shale.

Clapp and Deiss (1931, p. 690–691) discussed the source of the Belt sediments:

"Further, the Greyson, Spokane, and Marsh are much thicker than Wolcott estimated and are more arenaceous than the equivalent formations in the central and western parts of the Belt terrane. The most cogent evidence in opposition to Wolcott's concept of a 'somewhat distant source of supply' is the actual Archean mass upon which the Neihart quartzite lies, and the conglomerate, which Wolcott himself figured, at the base of the Greyson formation."

"... the Greyson formation seems to have been deposited only in the area of the Belt Mountains, inasmuch as nothing similar to it in lithologic character has been found in any other part of Montana, with the exception of the area 3 miles north of the town of Johns, on Prickly Pear Creek."

Farther west in the central and western portions of the Belt terrane the Greyson is absent

TABLE 1.—GENERALIZED STRATIGRAPHIC SECTION

Age	Stratigraphic Unit	General Lithology	Thickness Feet
QUATERNARY		Valley fill, alluvium, gravel fans, outwash and morainal material	0–300 +
	unconf.		
TERTIARY	Undifferentiated	Ash, silt, sand, and fanglomerate	0– ?
	unconf.		
	—unconf. ?—	Coarse conglomerate, some andesitic sand lenses Andesitic sandstone, sporadic conglomerate beds Coarse conglomerate, some andesitic sandstone	
?	Livingston fm.	Siltstone, shale, some andesitic sandstones and fresh-water limestones Andesitic sandstone, andesite conglomerates	up to 14,500
CRETACEOUS UPPER	Eagle sandstone	Salt and pepper sandstone, partly marine	100–600
	Colorado fm.	Black shale, rusty gray-green sandstone and siltstone, minor gray, salt and pepper sandstone Marine	1200–2400
? LOWER	Kootenai fm.	Basal conglomeratic sandstone and upper sandstone medial red-purple claystone and shale	386–447
	disconf.		
JURASSIC	Morrison fm.	Variegated shale and mudstone, with interbedded rusty calcareous sandstones	110–444
	ELLIS GROUP: Swift sandstone	Yellow calcareous sandstone, basal conglomerate or pebbly zone	50–100
	disconf.		
	ELLIS GROUP: Rierdon fm.	Massive grey, oölitic limestone, with overlying shaly beds	0–114
	ELLIS GROUP: Sawtooth fm.	Fine-grained dark-gray limestone, interbedded shale. Fragmental limestone with chert pebbles in lower part. Locally at top a red-yellow siltstone	20–145
	disconf.		
JURASSIC (?)		At many places a chert breccia and conglomerate occurs at this horizon	0–26
	disconf.		
PENNSYLVANIAN	Quadrant fm.	Pale-yellow to white, pure quartz sandstone or quartzite, locally calcareous. A few thin, light-gray dolomite beds	40–165
	Amsden fm.	Upper light-gray dolomite with some thin quartz sandstone beds	113–185
		Lower red siltstone unit with some variegated dolomite and impure fossiliferous limestone beds	11–189
?	disconf.		
MISSISSIPPIAN UPPER	BIG SNOWY GROUP	Cherty limestone Black shaly limestone and black shale	0–171
	local disconf.		
	BIG SNOWY GROUP	Locally red silty beds at top Red and yellow sandstone or calcarenite	0–163
	BIG SNOWY GROUP	Red siltstone, with a few purple and pale-yellow splotched dolomites Basal dolomite and dolomite or limestone breccia	0–100

TABLE 1.—*Continued*

Age	Stratigraphic Unit	General Lithology	Thickness Feet
MISSISSIPPIAN — LOWER	MADISON GROUP —disconf.— Mission Canyon fm.	Solution channels, caves at top Massive, light-gray limestone, solution breccias at several horizons	430–950
	Lodgepole fm.	Thin-bedded, yellow to red-stained, fossiliferous limestone Lower dark-gray, thin-bedded, less fossiliferous limestone	750–810
? DEVONIAN (?)	Sappington fm.	2–3 feet black silty shale Yellow sandstone grading downward into silty and sandy limestone Basal black shale	46–99
? DEVONIAN	—disconf.— Three Forks fm.	Gray limestone at top. Yellow siltstone and green shale. Middle ledge-forming, brecciated limestone. Basal evaporite solution breccias, and red and orange nodular limonitic shale	155–156
	Jefferson fm.	Light- and dark-brown, thick-bedded dolomite, dolomitic limestone, and limestone. A few intercalated yellow and pale pink dolomitic siltstone beds.	497–620
	Maywood (?) fm. — disconf. — —	Yellow mudstone and siltstone, thin dolomite beds. Grades upward into Jefferson. Red, blocky siltstone, with red-stained brecciated limestone beds in lower part. Basal red, fissile shale.	39–92
? CAMBRIAN — UPPER	SNOWY RANGE FORMATION — Sage pebble conglomerate member	Fine-grained, thin-bedded, dense limestone and limestone pebble conglomerate, with interbedded green shale. Basal columnar limestone.	121–204
	Dry Creek shale member	Gray-green fissile shale, with interbedded yellow calcareous siltstone and sandstone	42–76
	Maurice fm.	Massive light- and dark-gray mottled, oölitic limestone. Local reefoid development at base Thin to thick-bedded, gray, edgewise and flat-pebble limestone conglomerate with interbedded green shale Basal ledge-forming massive, oölitic, mottled limestone	363–433
MIDDLE	Park fm.	Green and maroon fissile shale, with thin-bedded limestone unit at top and, locally, intercalated conglomeratic, arkosic limestone and arkose beds in lower portion	188–192
	Meagher fm.	Thin-bedded, dark-gray, dense limestone with interbedded green shale Middle massive dark-gray dense limestone Thin-bedded, dark-gray, dense limestone with interbedded green and yellow silty shale	368–370
	Wolsey fm.	Green and maroon, micaceous shale with interbedded micaceous sandstone and siltstone. Locally contains conglomeratic arkosic limestone and arkose	152–210
	Flathead fm.	Red, pale orange, and white sandstone, locally quartzitic. Locally contains much feldspar, becoming arkosic. Conglomeratic in places	119–142

TABLE 1.—*Concluded*

Age		Stratigraphic Unit	General Lithology	Thickness Feet
		—unconf.—		
PRECAMBRIAN	ALGONKIAN	Belt series	Coarse, massive, poorly bedded arkoses and conglomeratic arkoses. Very coarse gneiss boulder conglomerate in southwest part of area. Interbedded dark-gray argillite and a few siliceous limestone beds in northern part of area	6000 +
		—Fault—		
	ARCHEAN (?)	Metamorphics	Gneiss, schist, metaquartzite, marble, injection gneiss, numerous pegmatite dikes, and veins.	?
			Approximate total thickness of section	27,000 +

clase, rarely containing small biotite books, occur in the metamorphic terrane. Locally injection gneisses are common; the invading material is granitic, and the host consists of various kinds of amphibole-bearing rocks.

The metamorphics are well banded, and foliation is easily measured. Flowage folds and other minor flexures are common. At normal stratigraphic contacts the attitude of metamorphic foliation and of bedding in the overlying Flathead sandstone is similar. West and southwest of Bridger Peak, the foliation curves around the noses of two plunging folds, nearly paralleling the overlying Cambrian strata. This suggests that the basement rocks have also been folded and faulted during more recent orogenic activity. The rock types in the metamorphic terrane suggest that most of this series was sedimentary, particularly the marbles, muscovite-rich quartzites, and garnet-rich amphibolites and gneisses.

In southwestern Montana two series of metamorphics, the Pony below and the Cherry Creek above, have been named and described (Tansley *et al.*, 1933, p. 8–12). Workers disagree on the validity of such a subdivision. Little is known of the relationship and ages of the metamorphics in southwestern Montana. For these reasons the term Archean (?) is used here.

Belt Series.—In the northern half of the Bridger Range more than 6000 feet of coarse clastics forms part of the normal Belt Series present north of this area. The rocks vary from extremely coarse conglomerates, composed of cobbles, boulders, and blocks of gneiss, quartzite, and schist, to coarse arkosic sandstones. The conglomerate matrix is about the same composition as the sandstones.

The sandstones contain much quartz, microcline, and orthoclase, with some plagioclase, plus a binding material of dark-green chlorite and altered biotite and amphibole. Invaginate outlines of quartz and feldspars, a feature inconsistent with sedimentary rounding, suggest slight alteration or metamorphism. (*See also* Merrill, 1893, p. 47–54.)

The rock weathers dark brownish green, and in places reddish. Fresh fractures are always moderately dark gray green. Much of the bedding is indistinct because the clastics are massive. Dips are difficult to ascertain because of this and because individual grains or particles are unoriented. Ripple marks, cross-bedding, mud cracks, and rain prints are also lacking, suggesting rapid deposition. Seemingly the water into which these clastics poured was much deeper than that of the Belt basin north of this area, and/or subsidence was rapid.

The coarse conglomerates are confined to the southwesternmost exposures; however, some gneiss pebbles and more quartz pebbles are common throughout the area. In the north beds of dark-gray argillite are interbedded with the arkoses, as are a few thin beds of siliceous limestone. In general coarse clastics diminish both to the north and to the east. This indicates a source to the southwest.

No contacts between the arkoses and the older metamorphic rocks were seen. The only visible contact is along the trace of Pass fault (Pl. 1). This fault marks the southern extent of

Beltian arkoses and the northern limit of Archean (?) metamorphics in the Bridger Range. This fault contact must be near the original depositional edge of the Beltian conglomerates.

Thesis, Princeton Univ., 1949) noted coarse arkoses interbedded with argillites or shales of type Belt appearance in the Horseshoe Hills area near Logan. He suggests that the coarse facies "interfingers laterally and grades

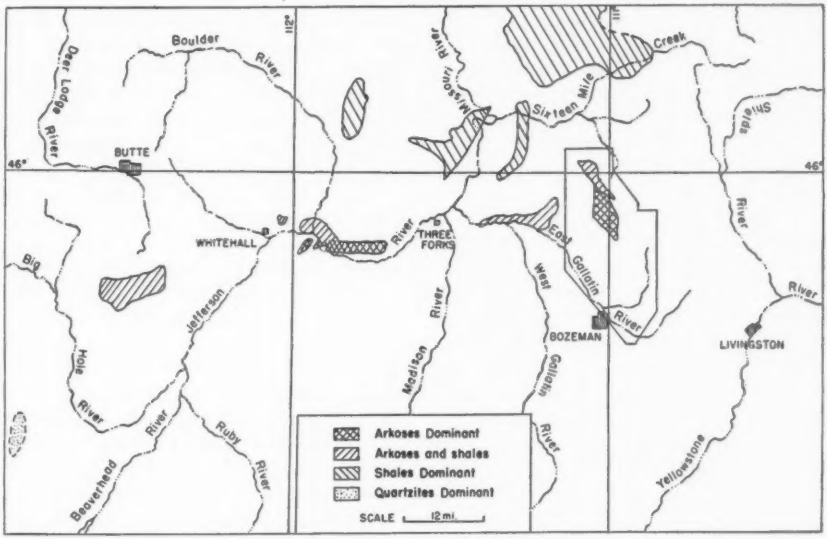


FIGURE 2.—LOCATION OF BELTIAN ARKOSES

Pass fault can be traced from Bill Smith Creek to Ross Pass and thence south-southeast across the headwaters of Maynard Creek on the east side of the Bridger Range (Pl. 1). Apparent displacement near the southeast end of the fault, along the dike, is less than 500 feet but increases gradually from Ross Pass westward to Dry Fork of Ross Creek where it is at least 1500 feet. Thence northwestward the displacement increases to about 5000 feet where the Flathead sandstone is in contact with the lower part of the thick Belt arkose sequence.

Farther northwest along the fault, the apparent stratigraphic displacement is opposite that described. The Laramide movement on this fault must have been opposite that along the same fault in Precambrian time (Fig. 5). Faulting must have controlled deposition of the coarse Beltian conglomerates. Nearly all workers have assigned the conglomerates a Beltian age. Recent studies have further clarified the relationships of these beds. Klemme (Unpub.

vertically" with the Spokane shale of the Belt Series. The base of the Belt Series is not exposed; deeper underlying portions, resting on the metamorphics, may be a shoreward facies of the next older formation, the Greyson shale.

Clapp and Deiss (1931, p. 690–691) discussed the source of the Belt sediments:

"Further, the Greyson, Spokane, and Marsh are much thicker than Wolcott estimated and are more arenaceous than the equivalent formations in the central and western parts of the Belt terrane. The most cogent evidence in opposition to Wolcott's concept of a 'somewhat distant source of supply' is the actual Archean mass upon which the Neihart quartzite lies, and the conglomerate, which Wolcott himself figured, at the base of the Greyson formation."

"... the Greyson formation seems to have been deposited only in the area of the Belt Mountains, inasmuch as nothing similar to it in lithologic character has been found in any other part of Montana, with the exception of the area 3 miles north of the town of Johns, on Prickly Pear Creek."

Farther west in the central and western portions of the Belt terrane the Greyson is absent

and the Newland (Lower Siyeh) limestone grades upward into the Spokane shales or argillites. Therefore, the Greyson must be a shoreward facies of part of the Spokane.

Alexander (Unpub. Thesis, Princeton Univ., 1951) found that the arkoses and arkosic conglomerates of the Jefferson Canyon area, east of Whitehall, interfinger with Greyson shales. In that area the beds are well exposed along the walls of Jefferson River canyon, and Alexander has named them the LaHood formation. South of the canyon, as in the Bridger Range, Middle Cambrian Flathead sandstone rests directly on Archean (?) gneisses and schists.

In the Rochester Mining District, 15–20 miles south of Butte, Sahinen (1938; 1950) noted coarse clastics interbedded with normal Belt sediments. A short distance to the south the Flathead sandstone rests with angular discordance on the Archean (?) metamorphics.

All the occurrences of the arkosic Belt facies fall on a nearly east-west line (Fig. 2). North of this line the coarse clastics grade rapidly into the normal Belt type of sediment; south of this line there are no rock units between the metamorphics and the Flathead sandstone. Obviously this line was a part of the Beltian shore line. The facts that the shore line was in general linear and that the clastics are so coarse along this zone and grade so rapidly to the north into dominantly argillaceous material suggest a fault along this part of the Beltian basin of deposition.

Deiss (1935) demonstrated the regional truncation of Belt formations over most of northwest and western Montana.

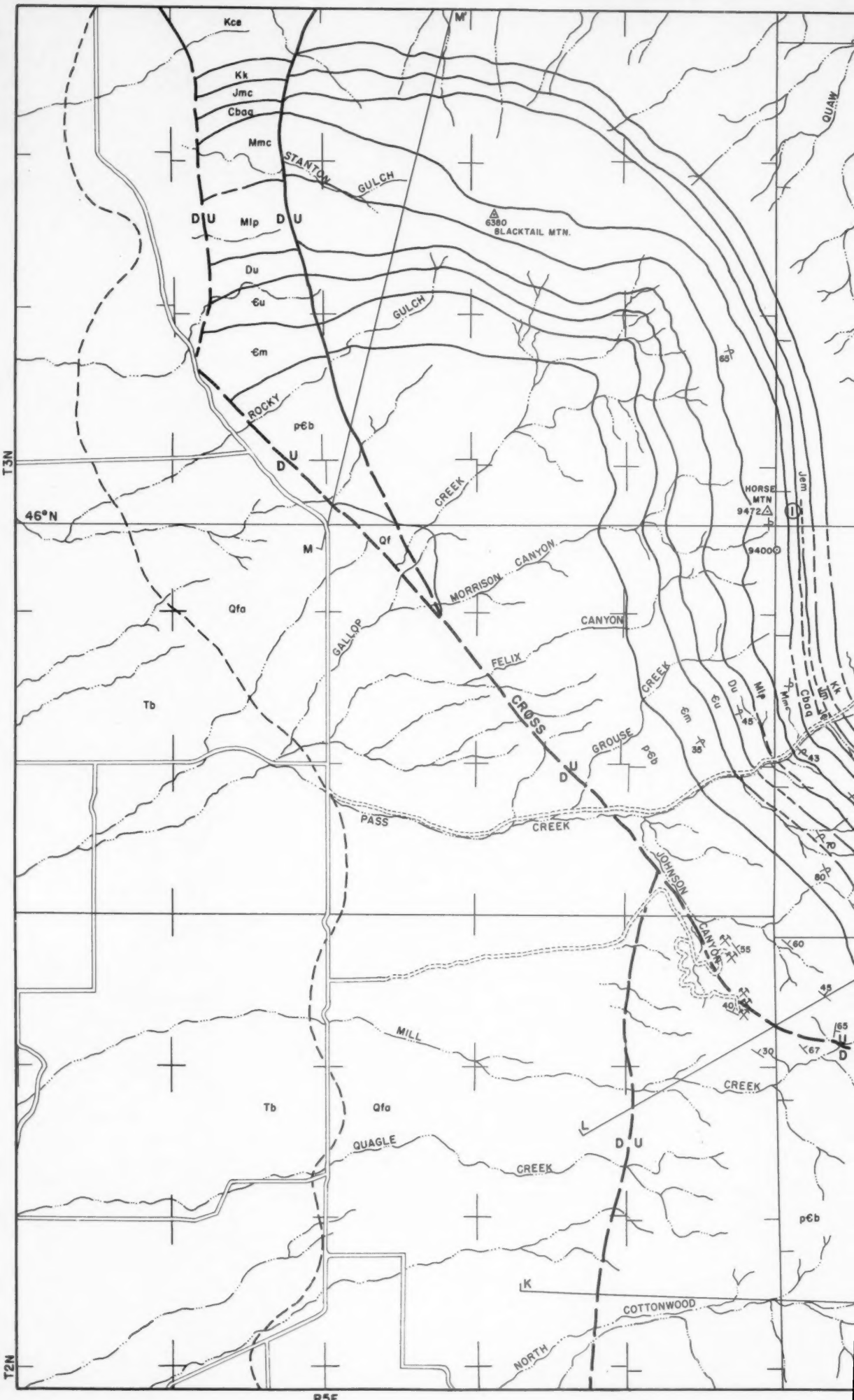
The wedge of coarse clastics that interfingers with Beltian rocks and that must have been derived from a rather pronounced positive area to the south suggests a long erosional interval between Beltian and Flathead deposition—*i.e.*, an Ep-Algonkian interval. The uniform widespread deposition of Flathead sandstone over both the Belt sediments to the north and the metamorphics to the south requires that the high-lying Beltian arkose source had to be reduced to a flat, featureless plain after Beltian and before Flathead deposition.

Cambrian

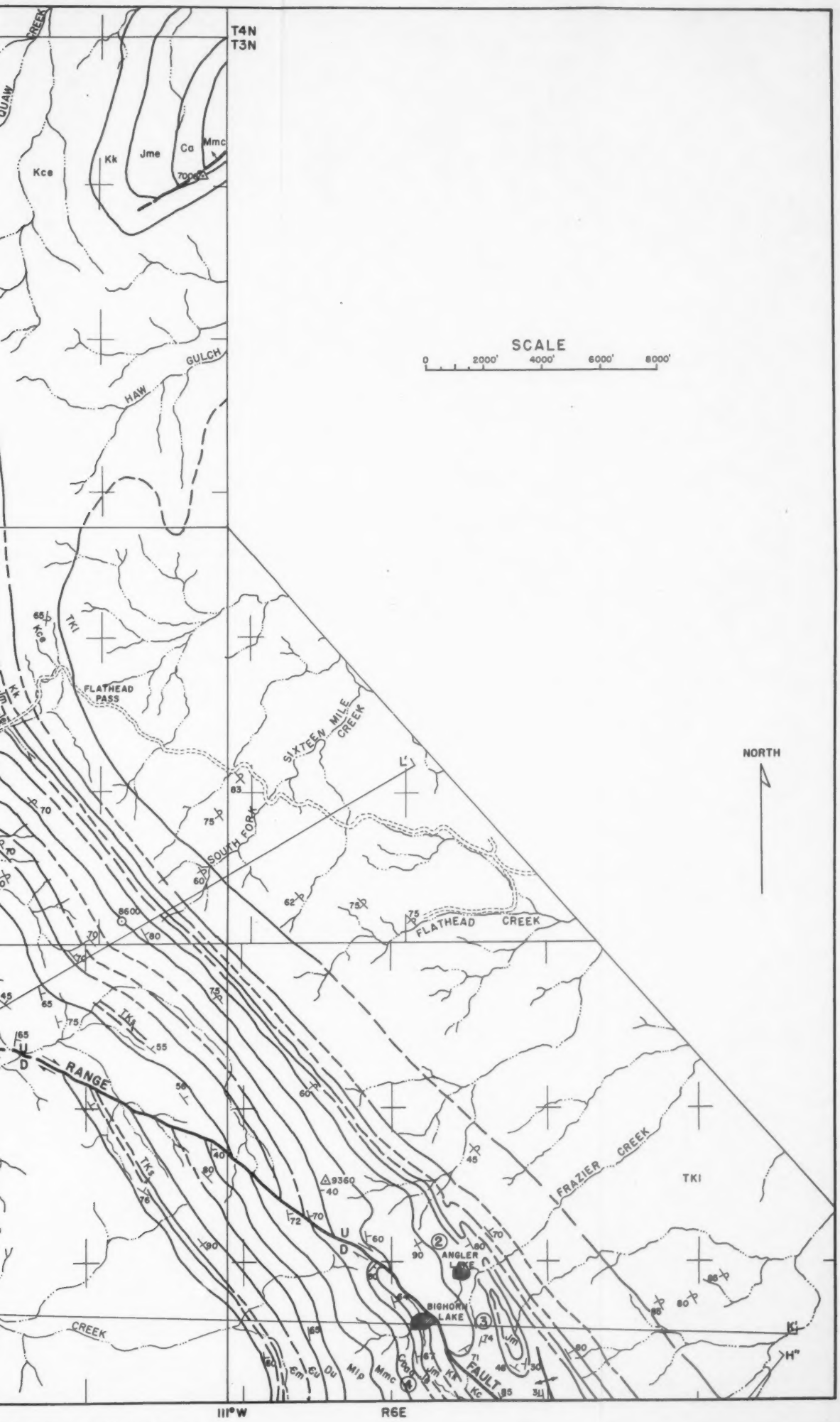
General.—Cambrian rocks are well exposed along the west side of the Bridger Range. Six formations are easily differentiated (from older to younger): Flathead, Wolsey, Meagher, Park, Maurice, and Snowy Range. The beds are 1560 feet thick in the northern portion of the map area, and in the southern part they total 1470 feet. Lithologically the section is much like the Cambrian section in the Horseshoe Hills area near Logan, Montana The Middle Cambrian units described by Deiss (1936, p. 1257–1342) and the Upper Cambrian lithologic divisions made by Lochman (1950, p. 2200–2222) can be readily distinguished in the Bridger Range.

The division of the local Upper Cambrian strata into the Maurice and Snowy Range formations follows usage by Lochman, whereas most workers refer to the same lithologic units as Pilgrim limestone and Dry Creek shale. However, as Lochman points out, the type Pilgrim of central Montana and the Pilgrim in southern and southwestern Montana are distinctly different. She also straightened out the usage of the term Dry Creek. As originally defined by Peale (1893, p. 24) the Dry Creek included only a thin shale zone overlying mottled limestones and underlying a series of limestone pebble conglomerates. Later workers misused the name, and as a result in most cases any beds above the mottled limestones and below definite Jefferson carbonates were referred to as Dry Creek shale.

In the area from the Horseshoe Hills and Bridger Range to the Beartooth Mountain region, the stratigraphic units of the Upper Cambrian correlate easily, both faunally and lithologically. For this reason Lochman (1950, p. 2205) suggests that the southern Montana formational names be used throughout the area—that is, Maurice formation for those beds previously called Pilgrim, and Snowy Range formation for the overlying shale and limestone-pebble conglomerate beds. She restricts the term Dry Creek to member rank, including only the beds originally called Dry Creek shale by Peale, and calls the overlying limestone pebble conglomerates the Sage pebble conglomerate member of the Snowy Range formation. The writer concurs in this usage. Lochman visited the Sacajawea Peak section during the summer of 1951; she believes the Cambrian rocks there are direct correlatives of those in the Horseshoe Hills and Beartooth Mountains.



GEOLOGIC MAP OF BRIDGER RA



R RANGE, MONTANA (North Part)

The Snowy Range formation has also been termed the Red Lion formation in the Horseshoe Hills area (Hanson, 1952, p. 17) because of the presence of ribboned limestone similar to that of the type Red Lion in the Philipsburg area. This type of limestone is not present in the Bridger Range.

Flathead formation.—In the Bridger Range the Flathead formation is fairly consistent in thickness (119–142 feet) but is variable in lithologic details. North of Ross Peak the formation consists of greenish and rusty, red, yellow, and white feldspathic sandstones and arkoses. These strata are thin- to medium-bedded, rarely cross-bedded, fine- to coarse-grained, variably glauconitic, and are sporadically conglomeratic, containing quartz pebbles. On the west side of Sacajawea Peak the formation is 142 feet thick. In the northern part of the range, north and east of Pass fault, the Flathead rests with apparent conformity on the Beltian arkoses; to the south it rests on pre-Beltian metamorphics (Pl. 1). West of Bridger Peak, in the southern part of the range, the Flathead consists of 119 feet of rusty-gray, greenish-white, tan, pink, and red, locally glauconitic and feldspathic sandstones. These sandstones are thin- to thick-bedded and sporadically conglomeratic. Cross-bedding and ripple marks are locally developed. These beds are less glauconitic and feldspathic than those of the northern portion. At nearly all exposures less resistant, platy, silty sandstone beds lie directly on the Precambrian rocks.

Although the formation rests with apparent conformity on the Beltian arkoses there is a marked disconformity at the base of the Flathead. The original type locality of the Flathead was near Flathead Pass in the northern part of the map area (Wilmarth, 1938, p. 738). In most of the area the term sandstone is more appropriate than quartzite. Deiss (1936) assigned a Middle Cambrian age to the Flathead sandstone.

Wolsey formation.—The Wolsey formation ranges from about 152 feet thick in the north to over 210 feet in the south. Gray-green to maroon, fissile, micaceous shale is interbedded with thin, rusty-brown, laminated, fine-grained glauconitic and micaceous sandstones. In the northern Bridger Range some thin glauconitic, arkosic limestone beds are also present, mainly in the upper part of the formation.

The formation rests with apparent conformity on the Flathead sandstone though the contact is sharp and marked by the abrupt change from sandstone to green fissile shale.

The Wolsey was originally called the Flathead shale; the type locality for the latter was near Flathead Pass in the northern part of the Bridger Range (Wilmarth, 1938, p. 738). The Flathead shale was renamed the Wolsey shale and given formational rank in 1900, by Weed, for exposures on Sheep Creek, near Wolsey, Montana (Wilmarth, 1938, p. 2362).

A few fossils are present in the Wolsey shale; in the middle of the formation a few phosphatic brachiopods and one trilobite, identified as *Zacanthoides* cf. *Z. cnopus* Wolcott, were found. The latter is common in the Wolsey and equivalent shales. Deiss (1936) assigned a Middle Cambrian age to this formation.

Meagher formation.—Throughout the Bridger Range the Meagher formation is persistently ledge forming. It is consistent in thickness (370 feet on the west flank of Sacajawea Peak and 368 feet on the west side of Bridger Peak) and can be divided into three units: (1) The lower 114–118 feet consists of light- to dark-gray, thin and irregularly bedded, fine-grained, dense limestone with intercalated greenish, calcareous shale. The shale layers are most prominent just above the basal 10–15-foot limestone unit and diminish upward. The upper beds of the lower unit consist of thin-bedded, dark-gray, dense limestone with thin, sporadic, yellow-brown, calcareous shale partings which locally produce a characteristic black and gold mottling (Hanson, 1952, p. 14). These beds are fossiliferous. (2) The middle 102–111 feet consists of ridge- or cliff-forming, dark-gray, massive, fine-grained, dense limestone. Bedding planes 1–2 inches apart have little effect on the weathering characteristics of the rock, except for a rough surface on massive outcrops. No fossils were found in these beds. (3) The upper 145–148 feet of the formation consists of light- to dark-gray, thin and irregularly bedded, nonresistant, fine-grained, fossiliferous, dense limestone with variable amounts of shale partings and beds.

The contact with the underlying Wolsey is obscure because the lower part of the Meagher is thin-bedded and nonresistant. In well-exposed sections the base of a moderately resistant,

10–15-foot limestone bed can be used as the contact.

Fossils identified by A. R. Palmer indicate a Middle Cambrian age: 5–10 feet below the top of the Meagher—*Glyphaspis* sp., *Ehmania* ? sp.; shaly limestone in the upper 40–50 feet of the Meagher—*Glyphaspis* sp.; 25–40 feet above the base of the Meagher—*Ehmania* sp. "Because of the confusion existing in the zonation of Middle Cambrian trilobites", Palmer (Personal communication, 1951) did not place this collection in any of the named Middle Cambrian zones. However, he considers them characteristic of younger Middle Cambrian rocks.

Additional fossils from the Meagher of the Bridger Range identified by B. F. Howell include: *Ehmania* cf. *E. wyomingensis*, from the lower 40 feet of the Meagher; and *Glyphaspis* cf. *G. perconcava* Poulsen, *Ehmania wolcotti* Resser, and *Arapahoia* sp., from the upper shaly member. According to Howell *et al.* (1944, Correlation Chart) these fossils indicate a medial Middle Cambrian age. However, in view of the paucity of work on Middle Cambrian zonation in the northern Rocky Mountains, it is best to assign the Meagher a medial Middle to late Middle Cambrian age.

Park formation.—The Park shale is marked by a topographic saddle between the underlying Meagher and the overlying Maurice limestones. It is remarkably consistent in thickness (188 feet at Sacajawea Peak, 192 feet at Bridger Peak). The formation consists mainly of gray, brown, green, and maroon, fissile, finely micaceous shale with some intercalated sandstone beds. In the southern part of the range the sandstones are thin, very hard, fine-grained, and calcareous, in many places containing abundant small phosphatic brachiopods. In the northern part of the range, especially in the area west of Sacajawea Peak, the Park shale has thin, intercalated beds of arkosic limestone, calcareous arkose, and arkosic intraformational conglomerate. These arkosic beds are most numerous in the lower part of the formation. Thin, intercalated limestone and shale beds characterize the top of the formation.

The contact of the Park with the underlying Meagher is arbitrarily placed at the top of the uppermost Meagher-type limestone, even though it includes some thin beds of Park-like shale.

A. R. Palmer (Personal communication, 1951) reports inarticulate brachiopods and undescribed trilobites of Middle Cambrian type from float near the Park-Maurice boundary. Calcareous sandstone beds in the upper third of the formation have yielded fossils identified by B. F. Howell as *Dicellomus occidentalis* Bell and *Homagnostus lochmanae* Howell and Duncan. Specimens from near the base of the Park shale were identified by Palmer as *Brachyaspidon* sp., *Ehmania* ? sp., and *Obulus* (*Westonia*) *ella* (Hall and Whitfield). These are late Middle Cambrian forms.

The lower ledge-forming limestone of the overlying Maurice formation yielded a specimen of *Bolaspidella wellsvillensis* (Lochman and Duncan). On the basis of this Palmer placed this bed in the lower subzone of the *Cedaria* zone. The position of the Park shale below this subzone supports a late Middle Cambrian age assignment.

Maurice formation.—In the Bridger Range the Maurice formation ranges from 433 feet thick at Sacajawea Peak to 363 feet at Bridger Peak. It contains two major divisions: (1) The lower 131–190 feet consists of a basal 8–10-foot gray-brown, massive, oölitic, glauconitic, vaguely mottled limestone, in places with a 6-inch bed of flat-pebble conglomerate at its base. This limestone yields fossils indicating earliest Dresbachian (early Late Cambrian) age. This ledge-forming bed is overlain by 123–180 feet of gray and yellow-brown, thin- to medium-bedded limestone, limestone pebble conglomerate, and edgewise conglomerate with interbedded gray-green, fissile, and calcareous shale. Shale is common in the lower strata and diminishes upward so that limestone-pebble conglomerates predominate near the top. These beds are fossiliferous. (2) The upper 232–243 feet of the Maurice consists of massive limestone. At Sacajawea Peak this unit consists almost entirely of medium- to thick-bedded, ledge-forming, dark- and light-gray mottled limestone (Pl. 3, fig. 1). The rock is fine- to medium-grained, oölitic, locally cross-bedded, and contains a few beds of flat-pebble conglomerate. At the top of the formation a 10–15-foot less resistant zone marks the contact with

the overlying Dry Creek shale member of the Snowy Range formation. The yellow and gray mottling of the latter zone differs from the mottling of the rest of the upper unit and weathers yellow brown. At Bridger Peak the lower 54 feet of the upper unit consists of a pale greenish-brown, massive, nonbedded, fine-grained limestone which weathers to a vague yellow, brown, and gray mottled appearance, and which locally has a columnar, reefoid limestone developed within it. The columnar limestone grades laterally and vertically into the nonbedded, massive portion of the unit. Above this the lithology of the unit is identical with that of the Sacajawea Peak section, consisting almost entirely of dark- and light-gray mottled limestone. Fossils are rare in this part of the Maurice.

The contact of the Maurice with the underlying Park is distinctive; it is at the base of a fairly prominent 8–10-foot, massive limestone. The basal ledge of the formation is early Late Cambrian in age, and fossils from the uppermost part of the formation indicate a late Dresbachian age (*Aphelaspis* zone). The age range for the Maurice of the Bridger Range is therefore the same as that of the Maurice in the type locality near Red Lodge, Montana (Howell and Lochman, 1936, p. 388–389).

A detailed study of the mottled limestones showed that the mottling is a product of partial dolomitization and recrystallization in irregular patches throughout the originally oölitic limestone (Pl. 3, fig. 2). The dolomitization was probably accomplished after deposition, but before or during solidification of the oölitic lime muds.

Snowy Range formation.—The Snowy Range formation is divisible into the Dry Creek shale member and the Sage pebble-conglomerate member. The latter gradually thickens from 121 feet near Sacajawea Peak to 204 feet southwest of Baldy Mountain. The Dry Creek thickness is less consistent and ranges from 50 feet at Sacajawea Peak to 42 feet at Bridger Peak and to 76 feet southwest of Baldy Mountain.

The Dry Creek shale member consists of gray-green, fissile shale with intercalated pale-orange to yellow-brown, calcareous sandstone and/or siltstone. In places there is a thin limestone pebble conglomerate at the base of the member, and there are also a few thin limestone pebble conglomerates in the upper 10–15 feet. The Dry Creek forms a conspicuous dark band and swale above the Maurice limestone on well-exposed slopes and is useful as a marker (Pl. 4, fig. 3).

The Sage pebble-conglomerate member has at its base a persistent 4–25-foot biostromal, columnar limestone, composed of the calcareous algae *Collenia magna* Fenton and Fenton. This unit is present at all exposures of the Upper Cambrian rocks except in the southernmost part of the range. Above the columnar limestone, the member consists of pale yellow-brown and pale-olive limestone and limestone pebble conglomerate. These strata are thin- to medium-bedded, fine- to coarse-grained, commonly glauconitic, and fossiliferous. They contain interbedded gray-green, fissile, calcareous shales. Characteristically the outer layers of the pebbles of the conglomerates in the upper part of the member are variegated greens, oranges, and purples. The well-rounded pebbles are invariably a fine-grained, dense limestone.

The contact of the Snowy Range with the underlying Maurice is marked by the abrupt change from massive limestone to shale, with no evidence of disconformity.

Fossils identified by A. R. Palmer indicate that the age of the Snowy Range formation ranges from early Franconian (medial Late Cambrian) to late Trempealeauian (late Late Cambrian). Lochman (1950, p. 2213) states that the columnar limestone at the base of the Sage pebble-conglomerate member is in the *Elvinia* zone of the standard Cambrian section. No diagnostic fossils were found in this unit in the Bridger Range, but its age is based on the stratigraphic identity and paleontological evidence from directly above the columnar limestone. The exact age of the Dry Creek shale is not known because no fossils have been found. It may include all or part of the uppermost *Aphelaspis* zone and the lower part of the *Elvinia* zone.

Cambrian-Devonian

Maywood (?) formation.—Above the Sage pebble conglomerate and below the Upper Devonian Jefferson formation lies a nonresistant,

dominantly siltstone sequence of variable thickness, herein called the Maywood (?) formation. The term Maywood was proposed by Lochman (1950, p. 2213) to include all beds above definite Cambrian and below definite Jefferson strata. The name comes from the Philipsburg quadrangle in western Montana, where beds of similar lithology and the same stratigraphic position were named the Maywood formation by Emmons and Calkins (1913, p. 7). The question mark is added here because of uncertainty as to age equivalence.

Northwest of Sacajawea Peak the formation aggregates 92 feet and consists of a basal 5-foot, red-brown, fissile, calcareous shale, overlain successively upward by: 1 foot of moderate reddish-orange, calcareous siltstone, containing sporadic limestone pebbles and glauconite grains; 45 feet of pale brownish-red, blocky-weathering, thin-bedded siltstone; 33 feet of pale yellow-gray, thin-bedded siltstone and mudstone; 6 feet of gray-brown, thin-bedded, platy, fine-grained limestone; and 2 feet of very pale orange-brown, massive, fine-grained limestone.

Farther north, on the rim of a large cirque at the head of Mill Creek, the basal fissile shale interval is about 30 feet thick. In the lower few feet of this shale are unidentified fragments of trilobite free cheeks and phosphatic brachiopods which B. F. Howell (Oral communication) considers Cambrian. Detailed search of the Maywood (?) throughout the rest of the area yielded no more fossils.

In the southern portion of the Bridger Range the lower red units of the Maywood (?) are infrequently present, and in most places the formation consists of alternating yellow to gray siltstone, dolomitic siltstone, and dolomite. The thickness ranges from 39 feet at Bridger Peak to 78 feet 1 mile farther west, to 75 feet southwest of Baldy Mountain.

The contact of the Maywood (?) with the underlying Snowy Range formation is marked in the northern part of the Bridger Range by a change from limestone pebble conglomerates to red, fissile, micaceous shale and red, blocky siltstones. In the southern part of the range the contact is less distinct and is placed at the change from limestone pebble conglomerates to yellow and gray dolomitic siltstones and dolomites.

At least the lower part of the formation in the northern half of the area is Cambrian; because of the gradational contact the upper part appears to be Devonian. The strata in the southern half of the range are lithologically equivalent to the upper part of the formation at Sacajawea Peak, and therefore Cambrian beds may not be present in the southern area.

Fossil evidence from stratigraphically equivalent beds in other parts of southwestern Montana indicate that the upper beds of the Maywood (?) formation may be Middle Devonian (Mann, Ph.D. Thesis, Princeton University, p. 18; Montis R. Klepper, 1951, Oral communication).

Since the formation is nonresistant and usually covered by debris from the overlying Jefferson formation, it has been mapped with the Upper Cambrian beds.

Devonian

General.—The Jefferson, Three Forks, and Sappington formations are discussed here, although there is some question as to the precise age of the Sappington formation. Throughout the Bridger Range the Jefferson overlies the Maywood (?) formation with a gradational contact. The three formations can be correlated lithologically with the Jefferson and Three Forks of the Horseshoe Hills area. In detail the Jefferson differs considerably from the well-known section at Logan, Montana. The Three Forks and Sappington formations, especially in the northern part of the Bridger Range, are identical with the Three Forks of the type region described by Haynes (1916b, p. 13–54).

A solution breccia and shale unit overlying the Jefferson dolomite is treated as a part of the Three Forks formation, rather than as an upper unit of the Jefferson, as suggested by Sloss and Laird (1947). This arrangement is logical for mapping, although it does not appear to conform with the original definition given by Peale (1893).

The Sappington sandstone, as defined by Berry (1943, p. 1–29) in the Three Forks area, is not a distinct mappable unit and is gradational with underlying strata. These beds have the same characteristics in the Bridger Range, and it is suggested that the Sappington be extended downward to include a series of cal-

careous siltstones and sandy limestones and an underlying, persistent black or gray shale horizon. At the top a black silty shale below the base of the Lodgepole limestone is also included in the formation.

Jefferson formation.—The Jefferson thins southward from 620 feet north of Sacajawea Peak to 497 feet at Bridger Peak. It consists of light- and dark-brown, medium- to thick-bedded, fine- and medium-grained dolomite, dolomitic limestone, and limestone of variable types, with intercalated pale-yellow and light-pink, thin-bedded dolomitic siltstone beds. The medium- and thick-bedded carbonates are characteristically rough-weathering, massive beds dominantly dark and granular, giving a simulated sandy texture on weathered surfaces; many are strongly petroliferous. Breccias are found at several horizons in the formation, as are beds containing small algal colonies. The thin-bedded dolomitic siltstone intervals are best developed in the upper part of the Jefferson. The uppermost bed of the Jefferson in the northern part of the map area is a pale-yellow calcarenite 5–15 feet thick.

The lower limestone and upper dolomite units of the Jefferson, described by Sloss and Laird (1947) in the Logan section, are not distinguishable in the Bridger Range.

Fossils are rare in this formation, and no systematic collections were made. Cooper *et al.* (1942) consider the Jefferson to be Finger Lakes and Chemung (early Late Devonian).

Three Forks and Sappington formations.—The strata between the Jefferson formation and Lower Mississippian Lodgepole formation is divided into the Three Forks and Sappington formations. This interval, except for a thin, black, silty shale at the base of the Lodgepole, has been called the Three Forks formation by most writers. For purposes of mapping the beds are here referred to as the Three Forks-Sappington.

In the vicinity of Sacajawea Peak the Three Forks is made up of (bottom to top): (1) greenish-yellow to red, hackly shale containing limonite nodules, and greenish-yellow to orange, massive, calcareous siltstone breccias, 85–95 feet; (2) yellow-brown, medium- to thick-bedded, massive, ledge-forming limestone locally brecciated and containing numerous calcite veinlets, 20 feet; (3) pale yellow-brown to gray, thin-bedded, calcareous siltstone or silty limestone which contains green shale in its lower and upper portions and which has a persistent gray limestone at the top, 45 feet.

In the same area the Sappington comprises (bottom to top): (1) black, fissile, conodont-bearing shale, 8–16 feet; (2) pale-brown to yellow-brown, thin- to medium-bedded, fine-grained, calcareous siltstone and sandstone noticeably more sandy in its upper part, 62 feet; (3) dark-brown to black, silty shale or siltstone, 2–3 feet. Cooper and Sloss (1943) call (3) the basal Lodgepole black shale.

Both black shales in the Sappington were described and tested for oil content by Condit (1919, p. 19):

> "Samples 390 and 391 were collected from the Three Forks formation at the west side of Ross Peak in the Bridger Range, in T. 2 N., R. 6 E., (should be T. 1 N., R. 6 E.) at an elevation of 7700 feet.... Results of the tests: Sample 390, oil, 1 gallon per ton. Sample 391, oil, 2 gallons per ton; nitrogen 0.22%, equal to 20.6 pounds of Ammonium sulphate per ton."

From his description of the section at this locality, it is obvious that sample 390 came from the black silty shale at the top of the Sappington, and sample 391 from the black shale at the base of the same formation. Both shales are mildly bituminous but are not of commercial importance in this area.

The Three Forks-Sappington interval thins slightly to the south and at Bridger Peak totals 205 feet. The details of lithology also vary somewhat. The basal shale-breccia unit of the Three Forks is replaced by a grayish-yellow, platy, calcareous shale; the ledge-forming, brecciated limestone is only 10 feet thick; the upper shaly part of the Three Forks thickens to 70 feet and is more calcareous. The basal shale of the Sappington is gray and silty in the vicinity of Bridger Peak, and the silty limestones and calcareous sandstones are only 39 feet thick.

The yellow-brown, calcareous sandstone and the upper black shale of the Sappington are conspicuous throughout the Bridger Range and are useful as reference beds in structurally complex areas.

A persistently fossiliferous, nodular siltstone bed just above the lower black shale of the Sappington contains *Syringothyris* sp. In a

section north of Bighorn Lake, 5–10 feet above this nodular unit, silty beds contain large spiriferids. Collections from these two zones were examined briefly by G. A. Cooper and A. L. Bowsher, and both (Oral communication, 1952) considered the two collections of Mississippian aspect. The fossiliferous nodular zone, containing a similar fauna, was also seen at Nixon Gulch, in the Horseshoe Hills.

Collections from the basal black shale of the Sappington, from a section north of Bighorn Lake, yielded conodonts which L. A. Thomas identified as species of *Hindeodella* and *Prioniodus*. He considered the types similar to those listed by Cooper and Sloss (1943) from the basal Lodgepole black shale (upper Sappington shale). Thomas (Personal communication, 1952) believes that the conodonts from both black shales have more affinity with Devonian types than with Mississippian. He states, however, that these beds may be transitional between the two systems.

REGIONAL RELATIONS: As originally defined by Peale (1893, p. 29) the Three Forks shales formation(included three units: a lower orange, reddish, and brownish zone of calcareous shales; a middle band of limestone 15–20 feet thick; and an upper zone of shales which included all beds up to the base of the Madison limestone. Haynes (1916b) discussed the Three Forks lithology and fauna of the type region in some detail and divided the formation into seven units in ascending order: a basal blacky orange and red shale, or limestone and shale, or locally a nodular limestone unit; a massive calcite-veined limestone; a green fissile shale which locally is a yellow to reddish shaly limestone in its lower portion; a gray limestone which weathers buff; a persistent black or dark-purplish shale; a buff-weathering, silty to sandy limestone which grades upward into the uppermost unit, a buff or yellow-brown calcareous sandstone.

Although the details for exact correlation of Peale's and Haynes' sections are lacking, the writer believes that Haynes has added two units to the lower part of Peale's section. Subsequent usage of the term Three Forks, in most instances, followed Haynes' definition. This usage is natural, for the formation is a mappable unit, whereas Peale's original definition would cause difficulties in mapping the basal contact. The seven units delimited in Haynes' Three Forks correspond closely to the lithologic characteristics in the Three Forks-Sappington of the Bridger Range area.

Berry (1943), while mapping in the Three Forks area, found a *Syringothyris* fauna near the base of a sandstone equivalent to the uppermost unit of Haynes' Three Forks. He named this the Sappington sandstone and assigned it a Mississippian age, yet recognized that the sandstone grades downward into the underlying silty and sandy limestones. However, he placed the latter beds, along with the subjacent black shale, in the Three Forks formation. Haynes (1916b) reported *Syringothyris* from as low in the section as 6 feet above the lower black shale. This suggests that the *Syringothyris* fauna is present in all the silty and sandy beds above the black shale and below the Lodgepole limestones. The writer collected that genus from the nodular silty limestone overlying the lower black shale, but it is absent from beds below that shale.

Sloss and Laird (1947, p. 1411) say, concerning Berry's Sappington sandstone:

> "However, it can be demonstrated that the normal shale facies of the Three Forks, bearing the typical *Cyrtospirifer* fauna, is interbedded with the sandy facies bearing *Syringothyris*, and that the sandy facies grades laterally into typical shale of the Three Forks formation."

The Sappington sandstone of Berry grades laterally and vertically downward into the silty and sandy limestone, but neither the Sappington nor the silty and sandy limestone grades into the typical Three Forks of Haynes—that is, into the gray limestone and green fissile shale of the type region. In all sections reported in the type Three Forks region, from Whitehall on the west to the Bridger Range on the east, a black to dark-purplish shale underlies the sandy and silty beds with sharp contact. It is difficult to understand how the gradation mentioned by Sloss and Laird is accomplished when the black shale always intervenes between the typical Three Forks lithology and the sandy-silty Sappington formation.

It has long been known that the fauna of the beds above the lower black shale is distinctly different from the fauna of the typical Three Forks shale (Peale, 1893; Raymond, 1907; Schuchert, 1910; Haynes, 1916b; Berry, 1943;

Cooper *et al.*, 1942). The gray limestone underlying the lower black shale bears a typical Three Forks fauna. Within 5–15 feet stratigraphically, a megafauna having a very different aspect appears. This is hardly a gradation in fauna, since this interval is occupied by a black shale containing neither megafauna.

The silty and sandy beds between the two black shales are a lithologically continuous unit characterized by a fauna of "Mississippian aspect" which contains conspicuous *Syringothyris*. The microfauna of the basal Sappington black shale is comparable to that of the basal Lodgepole black shale of Cooper and Sloss. This strongly suggests that the strata from the base of the lower black shale to the top of the upper black shale are closely related. The paleontological data apparently are not yet sufficiently diagnostic to permit assignment of either a Mississippian or a Devonian age to these beds. However, the faunas suggest a closer affinity to the overlying Lodgepole fauna than to the underlying Three Forks fauna. Possibly these beds transcend the systemic boundary, in which case they should be recognized as Devonian-Mississippian.

The presence of the black shale and the lithologic evidence of a changing environment, coupled with the faunal evidence, strongly indicates a distinct, though limited break in the depositional cycle at the bottom of the basal Sappington black shale. This conforms with the hiatus shown between the Three Forks and "Three Forks with *Syringothyris*" in the 1942 Devonian Correlation Chart (Cooper *et al.*, 1942).

The writer proposes[1] that the limits of the Sappington formation of Berry (1943, p. 14–16) be extended downward to include the underlying silty and sandy limestone beds and the lower black shale, and upward to include the black shale formerly called basal Lodgepole black shale by Cooper and Sloss. The type locality of the formation is considered that originally defined by Berry.

[1] In a paper which came out after this manuscript had been completed, F. D. Holland, Jr., appears to have come to the same conclusions concerning the Sappington formation, except that he still excludes the upper black shale from his Sappington formation (Holland, 1952, p. 1697–1734).

Mississippian

General.—The prominent Lower Mississippian limestones overlying the Sappington formation were formerly known as the Madison formation. The type locality was in the vicinity of the three forks of the Missouri River, the name being from the Madison River. In the Little Rocky Mountains of north-central Montana the formation was raised to group rank by Collier and Cathcart (1922, p. 173), who subdivided it into a lower Lodgepole formation and an upper Mission Canyon formation.

Sloss and Hamblin (1942, p. 315) extended the group rank and the formation names into southwestern Montana; because no specific type locality for the Madison had been designated by Peale, they chose the exposures along the Gallatin River, at Logan, as the type section for their Madison group. They further recognized subdivisions of the Lodgepole limestones which they called the Paine and Woodhurst members after terminology used by Weed (1900) in the Castle Mountains. Sloss and Moritz (1951, p. 2156) state:

> "The Woodhurst member is transitional above the Paine. Mappable contacts to separate the two members may be applied locally by using the lowest significant massive limestone bed, but it is apparent that the same contact can not be identified with assurance at widely separated exposures."

The writer recognized lithologic differences between the lower and upper parts of the Lodgepole formation in the Bridger Range but did not apply formal names to those units. The two formations, Lodgepole and Mission Canyon, are recognized and mapped.

The Madison limestones are prominent in the map area, since the crest of the range, with minor exceptions, consists of either the Lodgepole or Mission Canyon formations (Pl. 4, fig. 2).

Lodgepole formation.—In the Bridger Range the Lodgepole formation consists of approximately 750–810 feet of limestone with intercalated shaly limestone partings. At the base, overlying the upper black shale of the Sappington formation with a sharp contact, is 310–325 feet of dark gray-brown, mainly thin-bedded and platy, fine-grained, brittle limestones. Intercalated, in varying amounts, are beds of

yellow-weathering shaly limestones. Chert nodules are common in the lower part of this interval, and crinoid columnals and plates are also characteristic. This unit is sparsely fossiliferous. Above these beds lies 440–480 feet of pale yellow-brown, orange-brown, gray-brown, and red-stained, thin-bedded, fine-grained limestones, alternating with medium- to thick-bedded, massive, saccharoidal limestones. The saccharoidal limestone becomes more prominent upward in the interval, and also the upper portions are characteristically red or pink weathering. These beds are exceptionally fossiliferous.

The contact of the Lodgepole with the overlying Mission Canyon formation is arbitrarily placed. It is marked by a gradation, through 25–30 feet, from dominantly red weathering, thin-bedded limestones to light-gray, massive limestones.

Weller *et al.* (1948, Mississippian Correlation Chart) consider the Lodgepole Kinderhookian and early Osagian.

Mission Canyon formation.—In the map area the Mission Canyon formation is less consistent in thickness than is the Lodgepole formation. At Sacajawea Peak, where the Mission Canyon is overlain by Big Snowy beds, two sections of the former measure 950 feet and 870 feet thick. At Bridger Peak, where the formation is overlain by a partial and considerably thinner sequence of Big Snowy beds, the Mission Canyon is 718 feet thick. At Flathead Pass, where there are few if any overlying Big Snowy beds, the Mission Canyon is only 430 feet thick (Sloss and Hamblin, 1942, p. 322). This suggests a correlation between the thickness of the Mission Canyon formation and the amount of overlying Big Snowy group.

The Mission Canyon consists of pale yellow-brown, massive, poorly bedded limestones which weather light gray and form cliffs and castellated ridges. Chert and limestone nodules are common in the middle and upper parts of the formation. The limestone nodules are easily mistaken for chert. They are brown-weathering, ellipsoidal to irregularly shaped, fine-grained, dense limestone masses up to 1 foot thick and 1–3 feet long, and more or less parallel the bedding. The nodules contain, as inclusions, pellets or fragments of limestone identical to the enclosing limestone beds.

Solution or collapse breccias are conspicuous in the Mission Canyon formation, and in both the Sacajawea Peak and Bridger Peak sections three breccias occupy nearly identical stratigraphic positions. North and south of Sacajawea Peak two are prominent and can be traced for 2–3 miles. The blocks and smaller fragments in the breccias appear identical to the overlying limestone beds. The cement is sometimes a bright-red calcareous clay but is most often a yellow to gray limestone. The facts that the breccias occur at nearly identical stratigraphic positions and persist in the Sacajawea Peak area suggest they are continuous through much of the Bridger Range.

Collapse breccias have been reported in the upper part of the Mission Canyon limestone

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PLATE 3.—CAMBRIAN MAURICE LIMESTONE

FIGURE 1.—MOTTLED UPPER MAURICE LIMESTONE OF BRIDGER RANGE
Light areas dolomitized

FIGURE 2.—PHOTOMICROGRAPH OF MOTTLED UPPER MAURICE LIMESTONE OF BRIDGER RANGE
Oölites destroyed in dolomitized part

PLATE 4.—PALEOZOIC EXPOSURES IN BRIDGER RANGE, MONTANA

FIGURE 1.—SACAJAWEA PEAK, LOOKING SOUTH
Mmc—Mission Canyon, Mlp—Lodgepole, Ds—Sappington, Dt—Three Forks, Dj—Jefferson

FIGURE 2.—CREST OF NORTH-CENTRAL PART OF BRIDGER RANGE, LOOKING SOUTH
Showing Mississippian limestones as backbone of range

FIGURE 3.—PALEOZOIC SEQUENCE ON WEST FLANK OF SACAJAWEA PEAK, BRIDGER RANGE, LOOKING SOUTHEAST
pЄb—Belt series, Є f—Flathead, Є w—Wolsey, Є me—Meagher, Є p—Park, Є ma—Maurice, Є sr—Snowy Range, Dm—Maywood (?), Dj—Jefferson, Dts—Three Forks and Sappington, Mlp—Lodgepole.

FIGURE 1

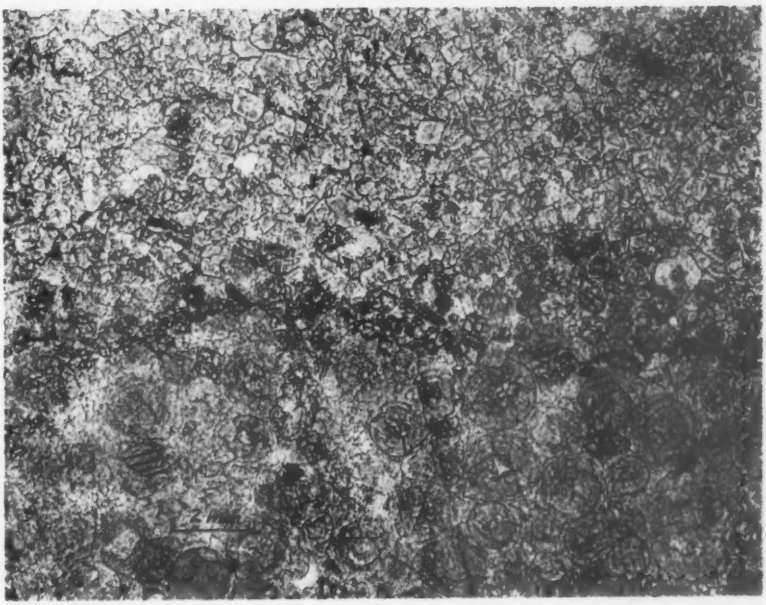

FIGURE 2

CAMBRIAN MAURICE LIMESTONE

FIGURE 1

FIGURE 2

FIGURE 3

PALEOZOIC EXPOSURES IN BRIDGER RANGE, MONTANA

FIGURE 1

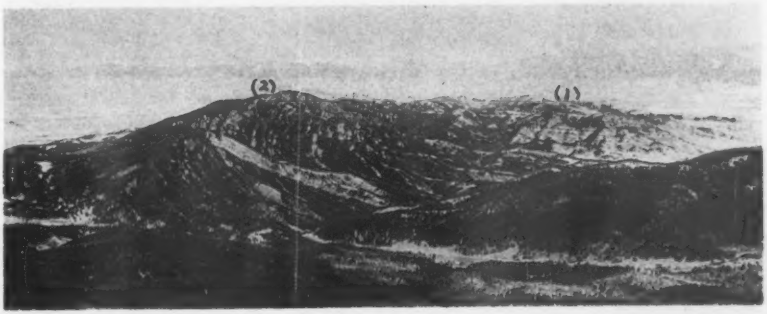

FIGURE 2

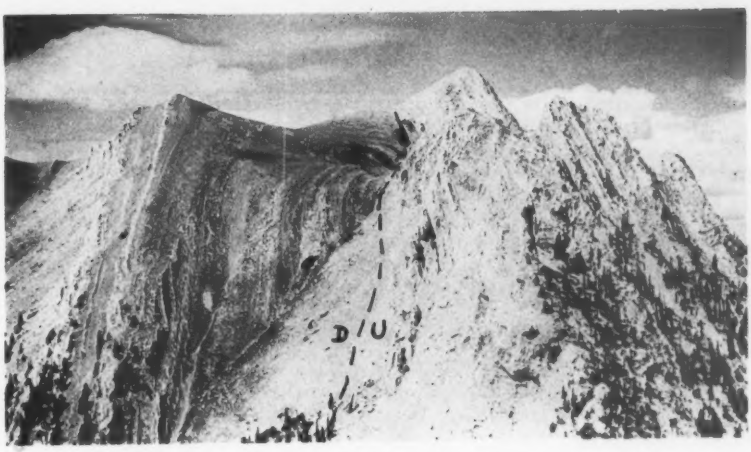

FIGURE 3

EROSIONAL AND STRUCTURAL FEATURES IN BRIDGER RANGE, MONTANA

FIGURE 1

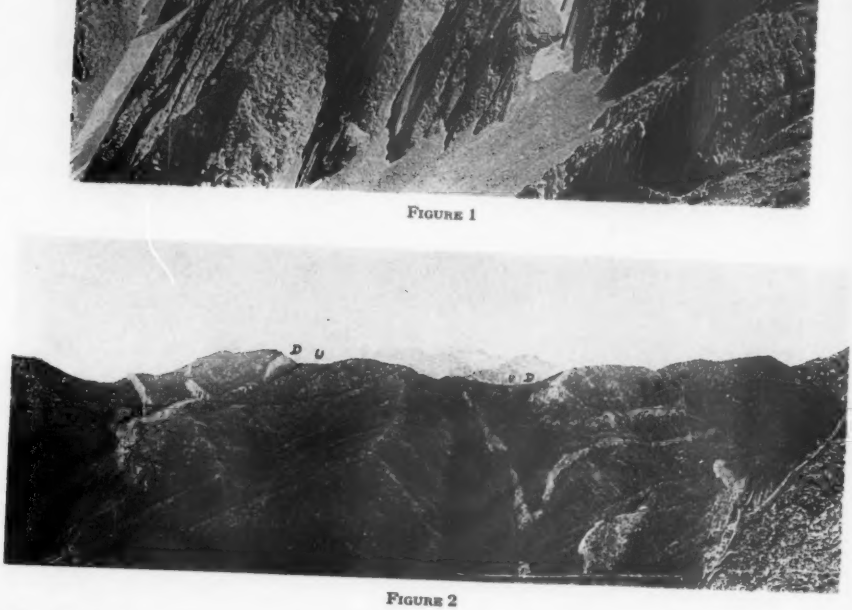

FIGURE 2

FIGURE 3

STRUCTURAL FEATURES IN BRIDGER RANGE, MONTANA

(Klemme, Unpublished Ph.D. Thesis, Princeton University, 1949; Klepper, Unpublished Ph.D. Thesis, Yale University, 1950). Klepper presents evidence that these stratigraphically localized breccias were formed by leaching at ground-water level in post-Madison, pre-Amsden time. His theory is that either a slight but uniform uplift or a widespread withdrawal of seas would have exposed a terrain of negligible relief. Under such conditions it is likely that ground-water level would have been constant, and leaching and ultimate collapse of the roof of the leached zone might have produced a rather continuous breccia zone at about the same horizon. The several persistent breccias in the Bridger Range thus would attest to comparative instability, in which ground-water level fluctuated. The fact that Big Snowy strata overlie the Mission Canyon in the Bridger Range suggests that the breccias developed prior to Big Snowy deposition, rather than prior to Amsden deposition.

Fossils are less conspicuous in the Mission Canyon than in the Lodgepole formation. Detailed collections were not made, however, and the writer follows Weller *et al.* (1948, Mississippian Correlation Chart) in considering the Mission Canyon formation Osagian.

Mississippian-Pennsylvanian

Big Snowy-Amsden strata.—Previous workers have not reported the Big Snowy group in the Bridger Range (Sloss and Hamblin, 1942; Gardner *et al.*, 1946); the diagnostic lithology and fossils are not conspicuous at the more accessible sections along the range. In much of the area the Big Snowy strata do not constitute a mappable unit and are therefore mapped with the Amsden and Quadrant. Twelve Big Snowy-Amsden sections were measured in an effort to establish their stratigraphic relationships (Fig. 3; Pls. 1, 2).

Five lithologic units have been chosen to depict the interplay of erosional, depositional, and tectonic factors involving this sequence of sediments. Within the Bridger Range these operational units, from older to younger, are:

(1) A basal red unit, 0–100 feet thick, overlying the Mission Canyon limestone, consists of red and pink, blocky-fracturing, dolomitic siltstone, and red to purple, hackly shale, with some thin, pale-gray or cream-colored, in places purple-mottled, fine-grained dolomite, and a few thin, red, calcareous sandstone beds. The basal portion in most places consists of a thin sequence of cream and lavender mottled, irregularly bedded, fine-grained dolomite, or a

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PLATE 5.—EROSIONAL AND STRUCTURAL FEATURES IN BRIDGER RANGE, MONTANA

FIGURE 1.—NORTH-NORTHEAST-SLOPING EROSION SURFACE CUT ON TERTIARY BEDS (MIDDLE DISTANCE) PASSES BENEATH PLEISTOCENE FANS TO LEFT
Looking northeast

FIGURE 2.—GENTLY DIPPING UPPER LIVINGSTON CONGLOMERATES (1) AND OVERTURNED LOWER LIVINGSTON CONGLOMERATES (2)
View northwest toward Battle Ridge

FIGURE 3.—CROSS RANGE FAULT BEDS TO LEFT OF FAULT DRAGGED DUE TO STRIKE-SLIP MOVEMENT
View northwest across Bighorn Lake canyon

PLATE 6.—STRUCTURAL FEATURES IN BRIDGER RANGE, MONTANA

FIGURE 1.—BEDDING-PLANE SLIP AND MINOR DRAG WITHIN LODGEPOLE FORMATION
Cross Range fault is beyond low saddle to right. View south toward Bighorn Lake.

FIGURE 2.—FOLDS AND FAULTS OF SCHAFER CREEK-MIDDLE COTTONWOOD SEGMENT OF BRIDGER RANGE
View north across Middle Cottonwood Creek

FIGURE 3.—FAULT-LINE SCARPS ALONG WEST FRONT OF BRIDGER RANGE
View southeast toward mouths of Bostwick and Middle Cottonwood creeks

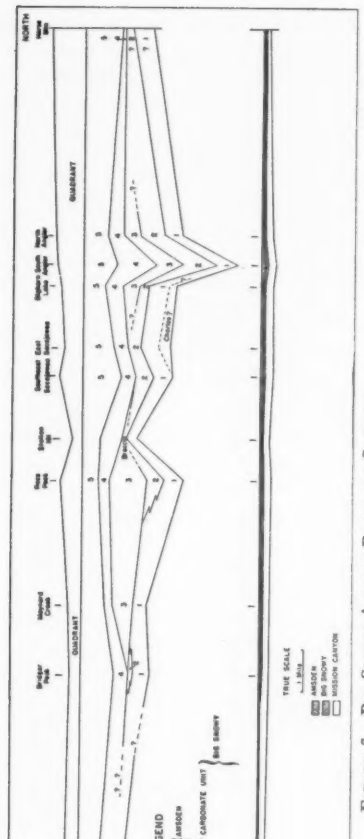


FIGURE 3.—BIG SNOWY-AMSDEN RESTORED SECTIONS

red-stained breccia of limestone and dolomite blocks and fragments.

(2) A sandstone or arenaceous unit, 0–163 feet thick, consists of red, pink, and pale yellow-orange sandstones, sandy siltstones, calcarenites, sandy dolomites, and some red siltstones. The composition varies considerably from section to section, but sandstones are most common. Both platy and massive beds are present. This unit is characterized by its quartzose sand content.

(3) This unit consists of persistently fossiliferous dark-gray to black shale, shaly limestone, and cherty limestone 0–171 feet thick. The lower shaly part grades upward into cherty limestone, and the gradational zone characteristically weathers yellow gray. The chert in the upper limestones is in most places black or dark brown. This is the uppermost unit of the Big Snowy group.

(4) The basal Amsden red unit, 11–189 feet thick, consists of red, dolomitic siltstones and some purple, hackly shales with thin interbedded, purplish, impure limestones which are fossiliferous in some places.

(5) The upper Amsden carbonate unit, 113–185 feet thick, consists dominantly of pale yellow-brown to white medium- to thick-bedded, fine-grained dolomites with some pale grayish-yellow to white, quartzose sandstones in the upper part. In some sections these interbedded sandstones are calcareous, in others quartzitic. A few limestones are found in the lower portion. The unit grades upward into the Quadrant formation. The contact is placed at the lowest thick sandstone or quartzite bed, where the section becomes dominantly sandstone. In the Bridger Range few thin dolomitic beds occur above this arbitrary contact.

Fossil evidence supports the assignment of the lower three units to the Big Snowy group and the upper two to the Amsden. J. B. Smedley (Personal communication, 1952) reports that collections from the black shaly limestone-cherty limestone unit are considered Chesterian. One collection from the basal part of the cherty limestone beds in the Bighorn Lake section, 255 feet below the Quadrant sandstone, yielded abundant *Leiorhynchus carboniferum* Girty which suggest correlation with the Heath formation of the Big Snowy group of

central Montana. A small collection from the top of the cherty limestone unit, in the same section, contains *Dictyoclostus inflatus* (McChesney) which suggests a Chester age. In the same section a prolifically fossiliferous zone in the upper part of the Amsden red unit, 102 feet below the Quadrant sandstone, is lower Pennsylvanian. On the southeast flank of Sacajawea Peak, beds 218 feet below the Quadrant sandstone, in the Amsden red unit, yielded fossils suggestive of both upper Mississippian and lower Pennsylvanian. The latter collection is stratigraphically lower than the Pennsylvanian fossils of the Bighorn Lake section.

Figure 3 illustrates the broad stratigraphic relations and variations in thickness of the five units along the Bridger Range, from Rocky Canyon, on the south, to Horse Mountain, north of Flathead Pass. A true-scale restoration of the Amsden, Big Snowy, and Mission Canyon illustrates the magnitude of the thickness changes. The geographic locations of the named sections are shown on Plates 1 and 2.

The strata of the Big Snowy group overlie the Mission Canyon from Bridger Peak northward to beyond the map area. The southern limit of the Big Snowy is unknown. No strata have been seen in the area south of Bridger Peak which could definitely be correlated with that group. It is difficult to distinguish the basal red Amsden from the basal red Big Snowy in the absence of intervening units. However, the writer regards as Amsden all the red beds overlying the Mission Canyon limestones in the southern portion of the Bridger Range and at Rocky Canyon.

The thickness of the basal Big Snowy red unit at the East Sacajawea and Bighorn Lake sections may be due to the presence of strata equivalent to the Charles formation of the central Montana subsurface. The basal carbonates and breccias in these sections are unlike any basal Big Snowy beds in the other sections, and they contain no typical Mission Canyon lithology.

The absence of the Big Snowy sandstone and cherty carbonate units at the Station 191 section (Fig. 3), as well as the absence of the carbonate unit at the Sacajawea Peak sections, is attributed to post-Big Snowy, pre-Amsden erosion. The sporadic distribution of the sandstone between Sacajawea Peak and Bridger Peak suggests either lenticular sand bodies or differential erosion in pre-black shale and cherty limestone time.

At the Ross Peak section, the black shales and shaly limestones of the Big Snowy are well developed and fossiliferous. Southward, black shale is absent, but the cherty limestone persists as far as Maynard Creek.

The preservation of the Big Snowy units in any particular section may be an indication of the positive or negative elements present during post-Big Snowy, pre-Amsden time. Thus Bridger Peak and the area to the south, as well as the area from Station 191 to Sacajawea Peak, appear to have been positive and subject to erosion. The Maynard Creek to Ross Peak area, and the Bighorn Lake to North Angler area were negative and preserved variously greater amounts of the Big Snowy units.

Where a fairly thick Big Snowy sequence was preserved, and especially where a Charles (?) equivalent is present, the Mission Canyon is thickest. Conversely, where the Big Snowy beds are thinner or absent the Mission Canyon is considerably thinner. This suggests that the areas which were positive in post-Mission Canyon (possibly post-Charles) time were also less depressed than adjacent areas during Big Snowy deposition.

The relatively thicker units in the Big Snowy and Amsden at the South Angler section suggest that this area was consistently more depressed. Further, this section is just northwest of a major cross-range fault. The inference is that this was a weak zone even in Big Snowy-Amsden time and that, with the onset of Laramide movements, the weak zone yielded by major faulting.

The relatively thick Big Snowy section at Ross Peak also lies near a concentrated zone of Laramide faulting (the Ross Pass fault zone) and may be similarly related to an old zone of weakness. This fault zone is directly related to the old Beltian shore line.

In summary, during Late Mississippian time the north and south extremities of Bridger Range were positive, while the central portion from Angler Lake to Bridger Peak was part of a mobile shallow trough in which various thicknesses of strata were deposited. It is not clear whether this trough was continuous with one

at Sappington Canyon, west of the map area, in which Big Snowy beds were deposited, or whether it was merely on a slight embayment along the margin of the main Big Snowy basin of deposition. The writer suggests that the Big Snowy strata of the Sappington Canyon, Three Forks, and the Lombard-Sixteen Mile Canyon areas were connected to the main Big Snowy basin of deposition across the area now occupied by the southern part of the Big Belt Mountains, north of the map area.

Pennsylvanian

Quadrant formation.—Throughout the Bridger Range the Quadrant formation is lithologically uniform but ranges from 50 to 165 feet thick. The thickness is least consistent in the central and northern parts of the range, the same general area of Big Snowy-Amsden instability. This may indicate continued differential movements in Pennsylvanian time, or the variations in thickness may have resulted from varying depths of erosion during Permian, Triassic, and Early Jurassic time, when this region was apparently above sea level.

The Quadrant formation is predominantly white, light-gray, and yellow to pink-hued, medium- to thick-bedded, in places cross-bedded, subrounded, medium- to fine-grained, pure quartzose sandstone, locally quartzitic. Commonly 2–4-foot massive, light-gray, fine-grained beds of dolomite occur in the lower portion of the formation. The beds are moderately resistant to weathering and in many places form a ledge or ridge.

At Horse Mountain, the formation is 165 feet thick in an exposure about 4 miles south-southwest of a section in which Klemme (Unpublished Ph.D. Thesis, Princeton University, 1949) reported only 15 feet of Quadrant. The formation is thinnest at Station 191 and at Bridger Peak (Fig. 3).

No fossils were found in the Quadrant formation of the Bridger Range. Thompson and Scott (1941, p. 349–353) and others assign a Pennsylvanian age to the formation.

Jurassic (?)

At many localities along the east flank of the Bridger Range 0–26 feet of chert breccia in a pale yellow-brown, calcareous sandstone matrix overlies the Quadrant formation and underlies the Jurassic Ellis group. Where 26 feet thick, on the southeast flank of Sacajawea Peak, the unit consists of the following, from base to top: an 8-foot chert breccia which is irregular in thickness along strike; a 2.5-foot chert and quartzite pebble conglomerate, which is slightly calcareous and irregularly bedded; a 2.5-foot nodular, irregularly layered chert bed; and a 13-foot yellow-brown to pale grayish-pink sandstone, which is medium-grained, slightly calcareous, irregularly bedded, and contains some chert and quartzite pebbles and fragments. These beds are directly overlain, in sharp contact, by a basal Ellis fragmental limestone made up of comminuted fossils, containing some chert pebbles and fragments.

At most localities the Jurassic (?) consists of variable thicknesses of chert breccia, and north of Fairy Lake it appears absent. No fossils have been found in the unit; its age, therefore, is not known. The abundance of chert in the breccias suggests it may be the product of erosion of the cherts of the Permian Phosphoria formation, which is known short distances west, northwest, and south of the Bridger Range.

At two localities in the Maynard Creek area the unit contains thin, variegated, calcareous shale zones. These shales were tested for phosphate but yielded negative results.

This breccia unit is here questionably assigned to the Jurassic and is considered a product of erosion of pre-existing Phosphoria formation cherts. It is mapped as part of the Ellis group.

Jurassic

Ellis group.—The type locality for the Ellis formation, as designated by Cobban *et al.* (1945, p. 451–453), is

> "along the north side of U. S. Highway 10 on the north side of Rocky Creek Canyon about 3.7 miles southeast of the site of Fort Ellis, or 7 miles southeast of Bozeman Court House, in sec. 19, T. 2 S., R. 7 E., Gallatin County",

in the southern extremity of the map area. This same section was described by Gardner *et al.* (1946, p. 22–23). At this locality the Ellis consists of three well-defined units which the

writer believes correspond roughly to the Sawtooth, Rierdon, and Swift formations delimited by Cobban (1945, p. 1262–1303) in the Sweetgrass Arch region to the north and which were applied throughout western Montana by Imlay *et al.* (1948). The uppermost formation, the Swift, is a yellowish-brown-weathering sandstone and sandy limestone with a basal chert-pebble conglomeratic zone. The Swift formation is 97 feet thick in a section measured by the writer on the crest of the Chestnut anticline, high on the north wall of Rocky Canyon. The Rierdon formation consists of a nonresistant, upper limestone and calcareous shale, about 30 feet thick, and a basal massive oölitic limestone. The Sawtooth formation, at this same locality, consists of an upper 30-foot, red to maroon shale, and a lower, 100-foot, gray shale and shaly siltstone unit which contains thin beds of dense, dark-gray limestone. This formation is exceptionally fossiliferous. At the base of the Sawtooth, on top of the Quadrant formation, there is a chert breccia which Cobban *et al.* (1945) include in the Ellis. Below this breccia occurs a nodular chert zone which the same authors refer to the Quadrant formation.

Though total thicknesses for the group are nearly the same in sections measured at Rocky Canyon, thicknesses of the Rierdon and Swift formations vary considerably. These variations are not unexpected though, since the basal conglomerate, regional relationships, and fossils of the Swift and Rierdon formations demonstrate a disconformity at the base of the Swift.

The writer believes the thin, red to maroon shale at the top of the Sawtooth formation is a westward-extending tongue of the upper part of the Piper formation of central Montana; this tongue thins northward and is not found north of the headwaters of Maynard Creek. Imlay *et al.* (1948) place this red unit and the underlying gray, shaly limestone of the Rocky Canyon section in the Piper formation. However, only the red unit is typical of the Piper lithology; the lower 100 feet is typical of the Sawtooth lithology of western Montana. Because of the dominance of Sawtooth lithology in this section, and because the red beds thin and disappear northward, the writer prefers to call the beds below the Rierdon the Sawtooth formation.

The gray, massive, oölitic limestone of the Rocky Canyon sections is a persistent marker bed throughout south-central and southwestern Montana (Imlay *et al.*, 1948). Locally in the Bridger Range, as in the Ross Pass section, and in a section near Fairy Lake, described by Gardner *et al.* (1946, p. 26), this oölitic member of the Rierdon may be thin, but normally it is persistently prominent. Where the oölitic unit is thin the overlying shaly limestone beds of the Rierdon formation thicken. The formations described are persistent throughout the Bridger Range, except at the northern extremity of the area, near Flathead Pass, where the Rierdon is cut out by pre-Swift erosion, and Swift conglomeratic sandstone rests on a thin remnant of the Sawtooth formation.

Thicknesses of the formations are variable. The Sawtooth ranges from 15 to 20 feet thick north of Flathead Pass, to a maximum of 149 feet in the section at Fairy Lake. One mile southward the Sawtooth is only 68 feet thick. At Ross Pass the formation attains another maximum of 145 feet, and southward it ranges from 88 to 130 feet thick. The Rierdon formation ranges in thickness from zero, north of Flathead Pass, to a maximum of 114 feet 1 mile southeast of Ross Pass. The formation thins abruptly south of the last locality to about 30 feet, and farther south, at Bridger and Rocky canyons, ranges from 60 to 78 feet thick. The Swift formation is about 100 feet thick in the northern portion of the Bridger Range, reaches minima of 39 feet at the Fairy Lake section and 50 feet at the Ross Pass section, and ranges from 60 to 90 feet thick in other sections in the range. The variations in thickness and in lithology probably result from instability of the depositional area. The maximum development of the Ellis group coincides with the Ross Pass fault zone, which was a mobile zone as far back as the late Precambrian.

Fossils are abundant in the Sawtooth formation and in the upper portion of the Rierdon formation. Locally, the Swift formation also contains numerous fossils. Systematic collections were not made by the writer, however, and he follows Imlay *et al.* (1948) in making

age assignments. Thus the Sawtooth is assigned to the Middle Jurassic, and the Rierdon and Swift to the Upper Jurassic.

Morrison formation.—In the Bridger Range the Morrison formation is less well exposed than the older strata, and relatively few sections were examined and measured. The formation consists of variegated red and green, blocky or hackly shales, siltstones, and mudstones, with intercalated yellow-brown to pale brownish-orange calcareous siltstones and standstone. In the thicker sections of the Morrison the upper 50–70 feet contain light- to dark-gray carbonaceous shales. Locally, the lower portion contains irregularly bedded impure limestones and limestone nodules.

At Rocky Canyon the formation is only 110 feet thick. To the north, at Bridger Canyon, it is 182 feet thick and on the headwaters of Maynard Creek it is 315 feet. The Morrison reaches its maximum thickness of 444 feet on the southeast flank of Sacajawea Peak. Northward it thins from 428 feet at Fairy Lake (Gardner *et al.*, 1946, p. 26) to 370 feet south of Bighorn Lake, and finally to 170 feet on the north flank of Blacktail Mountain.

No fossils were found in the Morrison formation of the Bridger Range. Imlay *et al.* (1948) consider the formation Kimmeridgian and early Portlandian or late Late Jurassic. The Morrison overlies the Swift sandstone conformably.

Lower Cretaceous

Kootenai formation.—The Kootenai formation is not well exposed in the map area, but its thick basal sandstone forms a topographic ridge in many localities. The formation may be divided into: (1) A thick, basal, white, gray, and pink-stained, massive, cross-bedded, coarse-grained, locally conglomeratic sandstone is a salt and pepper mixture of black chert and quartz grains, the latter having many euhedral overgrowths which lend a sparkly appearance to fresh and weathered surfaces. Locally, the sandstone contains numerous wood fragments. The upper part is a fine-grained, pure quartz sandstone, quartzitic in places. (2) The middle portion consists of variegated red, purple, and maroon, hackly shales and blocky mudstones, interbedded with rusty-gray to red-stained, impure, argillaceous sandstones. A few impure, silty, fresh-water limestones are present in the lower part. Nothing resembling the distinctive gastropod limestone of other western Montana exposures was found in the Bridger Range section. (3) The upper portion consists of medium- to thick-bedded, pure quartzose sandstone which contains a few thin, gray, silty, fissile, and platy shale beds. The sandstone, commonly iron-stained, consists of fine, well-rounded to subangular quartz grains. It is commonly quartzitic.

Two sections were measured in the map area. At Bridger Canyon the total thickness of the Kootenai was 386 feet, and the basal sandstone 55 feet thick. On the ridge south of Bighorn Lake, the total thickness was 447 feet, and the basal sandstone 82 feet thick.

No identifiable fossils were found in the formation. Previous authors consider the formation Lower Cretaceous. The contact of the Kootenai with the underlying Morrison formation is irregular and channeled in many places. The lower portion of the basal sandstone unit is conglomeratic. These features suggest a disconformable relation between the Kootenai and Morrison formations. The variability in thickness of the underlying Morrison, and the fact that where Morrison sections are thinner the upper part of the formation is absent, likewise indicate erosion between Morrison and Kootenai deposition. The contact of the Kootenai with the overlying Colorado formation is sharp and is marked by a rapid change from nonmarine quartzose sandstone to marine, dark-gray shale or silty sandstone.

Upper Cretaceous

Colorado formation.—The Colorado formation is a poorly exposed, nonresistant shale. No section suitable for measurement was found, and the total thicknesses given in Table 1 are calculated from the width of outcrop on topographic sheets. The upper contact with the Eagle formation is difficult to locate and is approximated in these calculations; the thicknesses of the Colorado and the Eagle are therefore only approximations.

In general the Colorado formation consists of dark-gray marine shales, iron-stained, gray-green, silty sandstones and siltstones, and,

locally, a few gray, thin-bedded, salt-and-pepper sandstones. In the lower-middle portion, at two widely separated localities, a quartzitic, black, chert-pebble conglomerate was found. A black chert conglomerate in the lower Colorado of the Castle Mountain area is persistent and marks an intraformational disconformity (Tanner, Unpublished Ph.D. Thesis, Princeton University, 1949).

No attempt is here made to subdivide the formation because of the lack of exposures. Cobban (1951, p. 2170–2198) demonstrated that the Colorado shale of central and northwestern Montana is divisible into equivalents of the several formations of the same age in the Black Hills region. The writer does not believe similar subdivision possible in the Bridger Range area, even with detailed studies.

Skeels (1939, p. 817) records 2550 feet for the Colorado-Telegraph Creek beds in the Trail Creek-Canyon Mountain area to the southeast. These beds are the stratigraphic equivalent of the Colorado formation of the Bridger Range. This compares favorably with the calculated thickness of 2400 feet for the Colorado near the southern end of the Bridger Range. The calculated thickness of 1200 feet in the Flathead Pass area does not compare well, however, with the 2800 feet of Colorado measured by Klemme (Unpublished Ph.D. Thesis, Princeton University, 1949) in the area north of the Bridger Range. The Eagle formation is also very thin in the Flathead Pass section, suggesting that this area subsided less during Colorado-Eagle time than did the areas to the north and south. Further, the Flathead Pass area was emergent or accumulated thinner sequences of sediments in post-Mission Canyon, pre-Amsden time, and also during Ellis and Morrison time.

Although described here under Upper Cretaceous the Colorado formation is considered uppermost Lower Cretaceous and lower Upper Cretaceous by most writers.

Eagle formation.—The Eagle formation forms prominent topographic highs along the southern and central Bridger Range. To the north, where it is very thin, however, it is inconspicuous and difficult to locate.

The basal contact is not well exposed, and the formation was therefore mapped with the Colorado formation. The contact with the overlying Livingston formation is gradational and not well exposed and is located only within limits of 100–200 feet at most places (Pls. 1, 2).

The Eagle formation consists of light-gray, thin- to thick-bedded, salt-and-pepper sandstone with intercalated carbonaceous shales and very thin coal seams. The lower portion commonly is greenish similar to parts of the underlying Colorado formation, and the upper part, similarly, is greenish much like the overlying Livingston formation. The sandstones consist dominantly of subangular grains of quartz, with some kaolinized feldspars and black chert, and are fine- to medium-grained, and well sorted. Locally, they are markedly cross-bedded.

Terrestrial, brackish, and marine fossils have been found in this formation (Stone and Calvert, 1910, p. 551–557). Thick-shelled pelecypods in the uppermost part of the formation, south of Cache Creek, indicate that at least part of the formation is marine as far west as the Bridger Range. The Eagle formation is considered early Montanan.

Upper Cretaceous and Paleocene

Livingston formation.—The Livingston formation (Weed, 1893, p. 21) is well exposed in the eastern quarter of the map area. No section was measured, but the thickness was calculated as at least 14,000 feet, east of the Bridger Range, on the basis of outcrop pattern and attitudes plotted on topograhic sheets. The Livingston thins markedly both to the north and south of this area of maximum thickness.

The formation can be divided into five parts: (1) A lower 4500-foot unit of tuffaceous or andesitic sandstones is composed of dark gray-green, medium- to thick-bedded, coarse-grained andesitic sandstones and andesitic conglomerates, with minor finer-grained, shaly beds and a few nodular limestone beds. The sandstones are composed of grains of quartz and plagioclase (andesine-labradorite range most common), moderate amounts of augite and chlorite, some biotite and magnetite, with small amounts of hornblende; in addition there are many small andesitic fragments. The conglomerates contain mainly well-rounded pebbles of andesite and a few pebbles of dark

quartzites, up to three quarters of an inch, averaging half an inch in diameter, in a matrix of andesitic sandstone. The basal beds are conformable and gradational with the underlying Eagle sandstone. (2) A finer-grained, valley-forming unit, about 3500 feet thick, consists of fine-grained andesitic sandstone, micaceous, hackly shales, and bentonitic clays, with some interbedded, coarse-grained andesitic sandstone. These beds are nonresistant and form the valley of the upper part of Bridger Creek. Colors are variable from several shades of purple to gray green, and pale greenish gray. The upper and lower contacts are gradational. (3) A conglomerate and andesitic sandstone about 3500 feet thick has coarse andesitic sandstones dominant in the lower 1500 feet. Coarse conglomerates gradually appear upward in the section, and the upper 2000 feet consists largely of pebbles, cobbles, and boulders of andesitic lavas with some olivine porphyrite, hornblende porphyrite, quartz diorite, and dacite (Weed, 1893, p. 10–41). There are also many pebbles, cobbles, and boulders of older sedimentary rocks including Colorado and Kootenai sandstones, Ellis sandstone and limestone, Quadrant quartzitic sandstone, Amsden dolomite, Big Snowy chert, Madison and Cambrian limestones, and Flathead sandstone, as well as Precambrian gneiss, schist, quartzite, pegmatite quartz, and granite. No material resembling either the Belt arkoses or the typical Belt sediments of the region north of the Bridger Range was found. The matrix is andesitic sandstone. Somber gray-green colors are predominant; these beds are resistant and form a steep, west-facing slope along Bridger Creek above the middle variegated shaly unit and form the backbone of Battle Ridge which trends northeastward to Wilsall on the Shields River. (4) An andesitic sandstone, about 2000 feet thick, contains sporadic conglomerate beds similar to those of the underlying unit. This sandstone does not crop out within the map area but occupies much of the western part of the Crazy Mountain basin south and southwest of Wilsall. The sandstones are strongly andesitic in the western part of its outcrop, but eastward, toward the Shields River, they are less andesitic as indicated by their lighter color and larger percentage of quartz. (5) An upper conglomerate unit, about 1000 feet thick, occurs as an erosional remnant preserved in a small area along the axis of a syncline which parallels the southeast side of Battle Ridge. These conglomerates consist of materials identical to those in the lower conglomerate unit—that is, a preponderance of andesitic lava cobbles and of older sedimentary, metamorphic, and igneous debris. These beds also do not contain detritus from the Belt series. The conglomerate matrix consists of andesitic sandstone; lenses and beds of andesitic sandstone are present, especially in the lower part, which appears to be gradational with the underlying sandstone.

The shaly, variegated, valley-forming unit was traced from Potter Creek, north of Wilsall, Montana, along the northwest side of Battle Ridge, into and through the upper valley of Bridger Creek, and on around the southwestern rim of the Crazy Mountain basin, into the type locality of the Livingston formation near the Yellowstone River (Pl. 7). This unit is distinctive lithologically and is notably more fossiliferous than other parts of the formation.

In a road cut in the lower part of the shaly unit at Bozeman Pass, 13 miles east of Bozeman, Montana, G. L. Jepsen[2] and the writer found Lancian vertebrate remains approximately 3200 feet above the base of the Livingston formation in the SW¼ NE¼ sec. 13, T.2 S., R.7 E., in the western part of the type area. About 3 miles to the east, at a road cut in the central part of sec. 16, T.2 S., R.8 E., two portions of a large dinosaur leg bone were found by members of the Montana State Highway Department. The exact spot from which these remains came is unknown because of the large-scale excavating methods used. However, the containing beds lie approximately 3700 feet above the base of the Livingston and in the lower portion of the variegated shaly unit. These large segments of limb bones also probably indicate a Lancian age.

Fragments of dinosaur bones were also found in exposures of the upper part of the shaly unit between Flathead Creek and Battle Ridge northeast of the map area. It is therefore

[2] G. L. Jepsen, of Princeton University, and a party composed of R. Witter, W. Morris, F. Goto, and R. Johnston.

apparent that most, if not all, of the shaly sequence is Upper Cretaceous and probably Lancian.

Plant specimens from a road cut 0.8 mile north of Wilsall in beds equivalent to the mid-part of the lower conglomerate series (unit 3) probably indicate a Paleocene age (Erling Dorf, Personal communication, 1952). A study of fresh-water Mollusca collected from various horizons in the Livingston has not yet been undertaken but may yield additional data concerning age relationships, especially concerning the upper two units.

It is apparent that the lower two units of the Livingston formation are Cretaceous. The exact position of the Cretaceous-Paleocene boundary is not known but is taken to be the boundary between the shaly unit and the overlying conglomeratic beds.

LIVINGSTON FORMATION AND LARAMIDE OROGENY: An extensive study of the literature has been made in an effort to establish a correlation between the Livingston as exposed east of the Bridger Range, the formation in the type area, and the Montanan, Lancian, and Paleocene sediments of the central Montana region. The results of this study have not been entirely satisfactory because of conflicting reports on type of lithology in certain areas, use of widely varying stratigraphic terminology, and erroneous identification and mapping of certain units in some areas. Tentative interpretations of the data, combined with field data, are given in graphic form in Plate 7.

Weed (1893, p. 21) named and described the Livingston formation from exposures near Livingston, Montana:

"Overlying the coal-bearing Laramie strata, there is a series of beds constituting a newly recognized formation, for which the name Livingston is proposed, as it is typically developed in the vicinity of Livingston, Montana. The formation consists of a series of beds, in places aggregating 7000 feet in thickness, composed of sandstones, grits, conglomerates, and clays, made up very largely of the debris of andesitic lavas, and other volcanic rocks, and including local intercalations of volcanic agglomerates."

He subdivided the formation into a lower, dominantly andesitic sandstone, called the leaf beds, and an upper purple shale and sandstone which graded upward into andesitic sandstone and conglomerate. Locally, in the Boulder River area, southeast of Livingston, the two major units are separated by a series of agglomerates. Weed believed what is now called Eagle sandstone to be Laramie (uppermost Cretaceous), and that the underlying beds were Coloradoan and Montanan. He mentioned the possibility that the upper part of the Livingston might be equivalent to the Fort Union and assigned the formation to the Cretaceous (?).

The units described by Weed correspond to the lower three units of the Livingston as delimited for the area east of the Bridger Range. The 2000-foot sandstone unit and the 1000-foot conglomerate unit of the latter area are not present in the vicinity of the type locality.

Although Weed's age assignments and some of his interpretations are in error, his descriptions of the occurrence of the strata in the field are quite lucid. Concerning the purple shaly unit he wrote (Weed, 1893, p. 30):

"Overlying the leaf beds and volcanic agglomerates there is a series of shales and interbedded sandstones and grits that form a prominent part of the formation wherever the upper portion is exposed. In the eastern part of the region the sandstones are fine and uniform in grain, making an excellent building stone in the vicinity of Big Timber. Traced westward these beds are easily recognized by their red soils that form the terrace flats to the south of the Yellowstone River as far as Springdale. Beyond here they form the low hills and the country north of Livingston. It is in these same shales that the valley of Billman creek has been cut, and they have also afforded the material out of which the long strike valley east of the Bridger Mountains has been eroded. . . . Northward the open valley of Flathead creek and the synclinal basin north of the Shields river valley are both eroded in this part of the Livingston formation. . . . This series of beds has thus far failed to yield any fossil remains."

Thus Weed clearly recognized the broad similarities between the lower units of the formation from the Boulder Creek area to the Bridger Creek area, and on to the north and northeast of the latter.

The purple shaly unit has been traced from near Livingston to the Bridger Creek area, and these beds have yielded vertebrate remains which indicate a Lancian age. Farther east, where the purple shale series overlies the agglomerate, or so-called igneous series, recent writers have called the shaly beds the Hell Creek formation or Lance (Parsons, 1942; Stow, 1946). Plant material from the overlying conglomeratic sequence indicates a probable Paleocene age for the containing beds. The

preceding data are significant in that they show that the major pulses of orogeny resulting in the coarse conglomerates were post-Lancian.

Stone and Calvert (1910, p. 551–557, 652–669), who worked extensively on the coal-bearing Eagle formation (Laramie of Weed) and the Livingston formation, presented data which did much to clarify the stratigraphic relations of these units. Their fossils clearly demonstrate that the Laramie of Weed (1893) is actually lower Montanan and correlative, at least in part, with the Eagle sandstone of central Montana. They also showed that Weed's Colorado-Montana beds below his Laramie were all Coloradoan in age, and that beds in the lower third of the Livingston formation, in the region northeast of the Bridger Range, were in part of Judith River and Bearpaw age. Their collection of Bearpaw marine invertebrates from 2500–3000 feet above the Eagle formation, a few miles north of Wilsall, proves that the Bearpaw sea was present in the area west of the Crazy Mountains. These authors further showed that the Livingston grades eastward, around the north end of the Crazy Mountains, into strata identified as Montanan and Paleocene (Lebo).

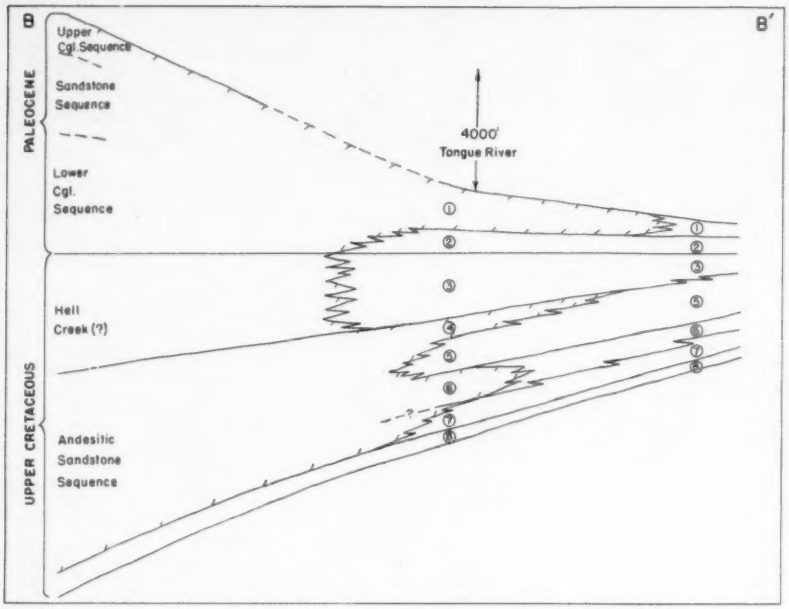
Klemme (Unpublished Ph.D. Thesis, Princeton University, 1949) described a 4000-foot series of purple to gray-green, coarse, andesitic detrital-lithic and crystal-lithic tuffs, volcanic conglomerates, and tuffaceous limestone in an isolated outcrop area (the Maudlow basin) directly north of the Bridger Range. About 300 feet above the Eagle sandstone at the town of Maudlow, on Sixteen Mile Creek, he found Lancian fossil plants. He also mapped a well-defined conglomerate (Maudlow conglomerate lentil) about 600 feet above the Eagle sandstone at the latter locality. The lentil is 688 feet thick and consists of andesitic volcanic conglomerate and conglomeratic, crystal-lithic tuff. In the upper part of this lentil, and 300 feet above it, Klemme found Paleocene plant remains. The absence of the thick sequence of Montanan beds, present just east of Klemme's area, suggests that the base of the Livingston rests disconformably on the Eagle in the Sixteen Mile Creek area. An interpretation of the relationships between the latter area and the area north of the Crazy Mountains is shown diagrammatically in restored section (Pl. 7, A-A').

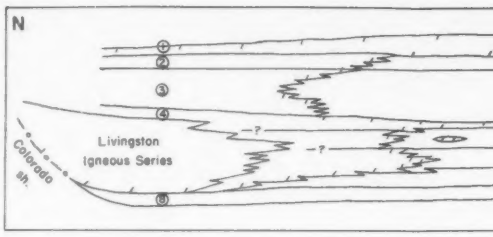
South of the Yellowstone River, along the north front of the Beartooth Mountains, a local igneous facies is intercalated with the lower part of the Livingston formation. Weed (1893), Vhay (1939), and Parsons (1942) have described these as breccias, tuff-breccias, lapilli-tuffs, tuffs, vent agglomerates, tuffaceous sandstones, and volcanic conglomerates. The igneous series is at least 2000 feet thick in the Deer Creeks' area and thins to 50 feet at the Yellowstone River, near Springdale, and also wedges out southeast of the Stillwater River, east of Nye, Montana. Paleobotanical evidence (Parsons, 1942) indicates that the first extrusive activity in this area occurred in Judith River time. Locally, uplift and erosion during extrusion resulted in unconformable relations between younger andesitic pyroclastics and older pyroclastics and sediments. Weed (1893) shows that, as the igneous series thins westward, the underlying tongue of Livingston sediments thickens. According to Parsons and Stow (1942, p. 344–345) this igneous member of the Livingston formation:

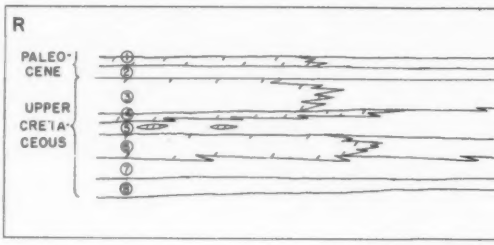
> "grades eastward, northward and westward into volcanic sandstones and conglomerates, and finally into Judith River, Bearpaw, Lennep, and Lance sediments. . . . The Hell Creek member of the Lance formation (now Hell Creek formation) overlies the volcanic sandstones and conglomerates with a conformable, gradational contact."

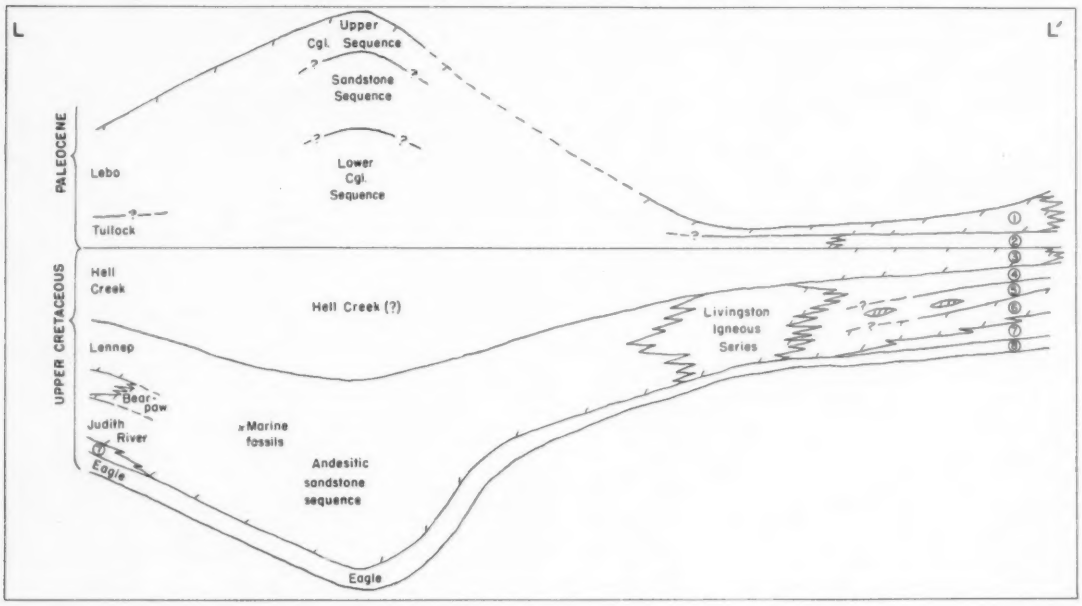
Other local sources for the igneous material of the Livingston formation are reported at Flathead Pass in the Bridger Range (Weed, 1893) and on the south fork of Smith River (Tanner, Unpublished Ph.D. Thesis, Princeton Univ., 1949).

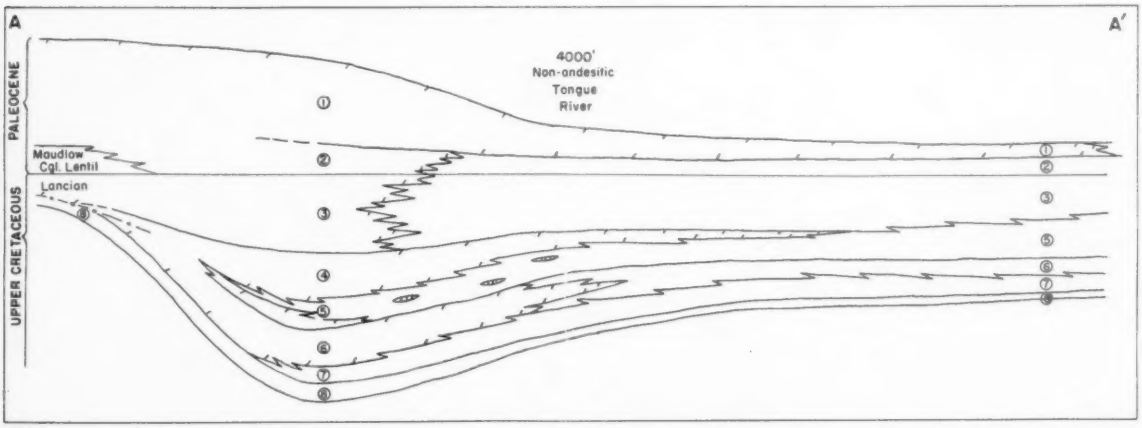
DISTRIBUTION OF ANDESITIC MATERIAL: The approximate limits of noticeably andesitic character of the Judith River, Lennep, Hell Creek, and Lebo formations are shown on Plate 7, which also shows the distribution of the Livingston igneous facies, as mapped by Parsons (1942) and Stow (1946). It also shows the base of the Livingston formation in the main part of the Crazy Mountain basin. The latter boundary is taken in part from Stow (1946), from Brown (1949), and from the 1944 Montana State Geologic Map. This contact is approximately at the top of the purple shaly



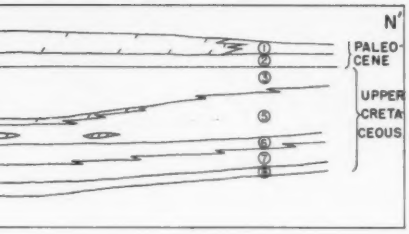


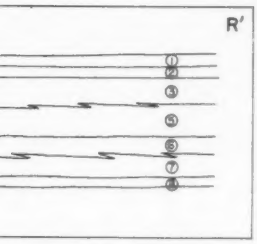


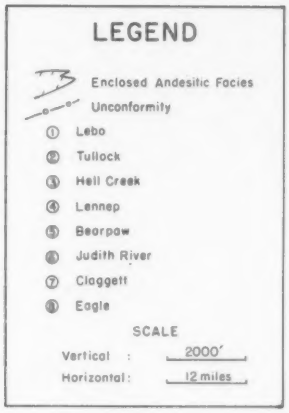


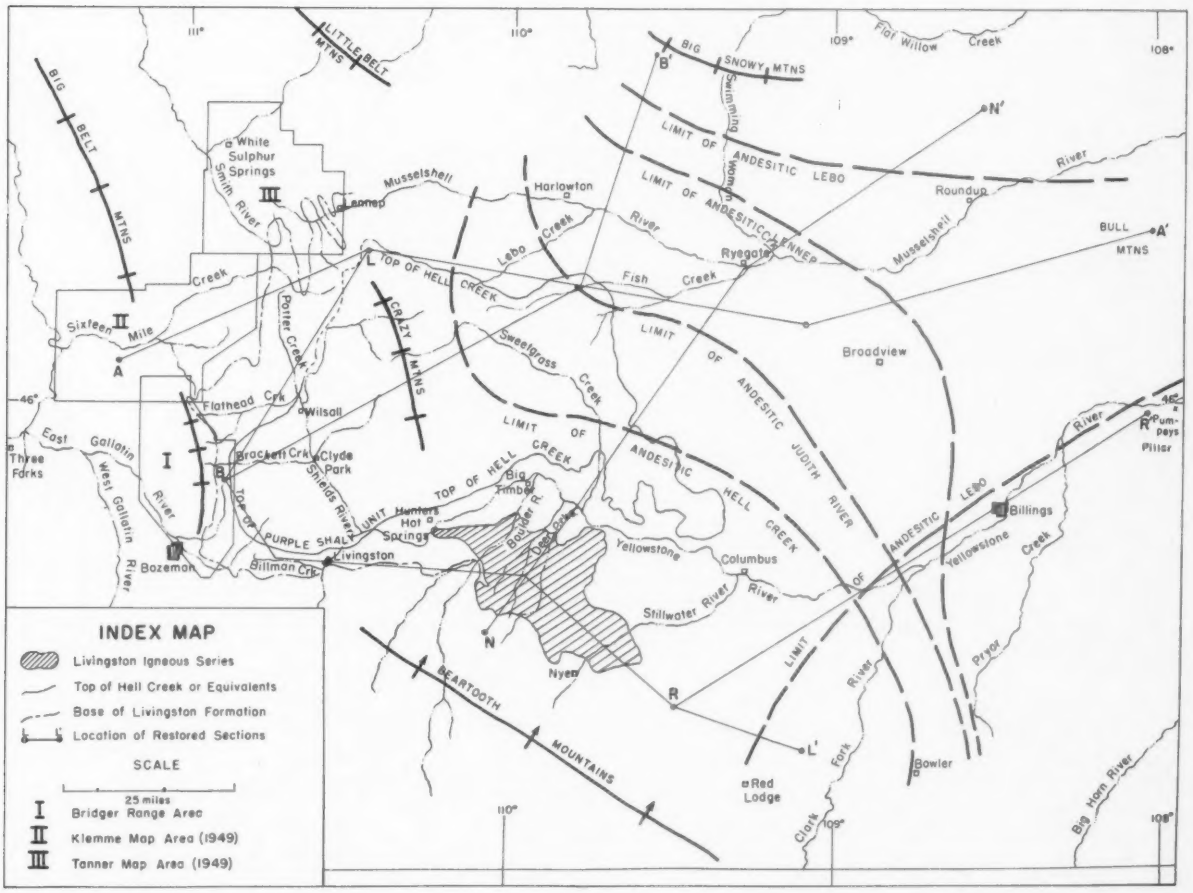


MAP AND RESTORED SECTIONS SHO
OF LIVINGSTON FORM









S SHOWING RELATIONSHIPS
FORMATION

unit of the Livingston formation in the Crazy Mountain basin between Hunter's Hot Springs and Wilsall. The position of the line is generalized from Wilsall to the north end of the Crazy Mountains.

Restored sections (Pl. 7) illustrate the relationships of the Livingston to the Upper Cretaceous and Paleocene sediments of the south-central Montana region. Three tongues of terrestrial andesitic sediments extend from the western and southern margins of the Crazy Mountain basin into the central Montana region: the Judith River, Lennep, and Lebo formations.

Eastward the Montanan sediments become marine. The Judith River tongue loses its andesitic character, although still terrestrial, and finally wedges out into the Pierre shale of southeastern Montana and the Dakotas. The Lennep tongue wedges out in central Montana so that terrestrial Hell Creek beds overlie the marine Bearpaw shale. Eastward across southern Montana the Lennep grades into brackish-water deposits and then into the marine Fox Hills sandstone of southeastern Montana and northeastern Wyoming. The Bearpaw shale represents the last extensive advance of the Late Cretaceous seas in the Montana region. Younger sediments in Montana and the Dakotas are terrestrial, except in a small area in North Dakota where the Cannonball sea persisted into Paleocene time. Within this vast sequence of terrestrial sediments the Lebo tongue is distinctive because of its andesitic character. Eastward beyond the limits of its andesitic character the Lebo is a terrestrial shale.

INTERPRETATION: According to Stone and Calvert (1910, p. 652–669) and Tanner (Unpublished Ph.D. Thesis, Princeton University, 1949) lenses of andesitic material in the Eagle sandstone from the northern part of the Bridger Range to the area west and northwest of the Crazy Mountains show that andesitic material was being supplied by early Montanan time. The strongly andesitic nature of beds overlying typical Eagle sandstone in the Bridger Range area and on the headwaters of the south fork of Smith River suggests that volcanic material was being supplied in considerable quantity during Claggett time. However, the limited extent of this material in the Claggett farther east indicates that volcanism was not at its maximum intensity in near-by source areas.

The local sources of volcanic materials reported by Parsons (1942, p. 1175–1186) and Tanner (Unpublished Ph.D. Thesis, Princeton Univ., 1949) were already active in Judith River time. These and other local sources near the margins of the Crazy Mountain basin and to the west could account for the advance of andesitic debris eastward and northeastward.

A combination of factors, possibly a lull in volcanic activity and uplift to the west with concomitant subsidence in the Crazy Mountain basin, allowed the Bearpaw sea to advance into the area west of the present Crazy Mountains. However, andesitic material was still being supplied to the basin in Bearpaw time, as evidenced by the continuous andesitic section in the type Livingston and Bridger Creek areas, by the interrelations of the Bearpaw formation and the Livingston igneous series, and by the andesitic lenses in the Bearpaw shale for a considerable distance into the central Montana area.

Between the Big Snowy Mountains and the Yellowstone River, the Lennep formation is distinctly andesitic to its furthermost limits. East and northeast of this the Hell Creek rests on the Bearpaw shale. The Lennep tongue probably reflects uplift and/or increased volcanism in the source areas to the west and southwest. The absence of beds of Montana age in the Sixteen Mile Canyon area, probably indicating a pre-Lancian unconformity, supports this conclusion.

Apparently during Hell Creek there was a lull in andesitic volcanic activity, and/or decreased erosive action in the source areas, because the materials deposited (the purple shaly unit of the Livingston) are noticeably finer-grained than the underlying and overlying strata. However, some andesitic sandstones and bentonitic clays and shales show that some volcanic debris was still being supplied to the depositional area.

Along the north front of the Beartooth Mountains and to the east, wherever the Tullock formation is recognized, the beds are nonandesitic, probably suggesting that volcanism was dormant at that time in the area south of the Crazy Mountain basin. However, in the

basin west of the Livingston igneous series the Tullock equivalent is andesitic and occurs in the coarse andesitic sandstones just above the purple shaly unit. The coarse andesitic detritus, and the progressively coarser sediments upward into the lower conglomerate sequence, suggests that strong deformation began in early Paleocene time, became progressively more intense, and culminated in deformation and uplift which produced the coarse material of the upper Livingston conglomerate units.

The Lebo is the most extensive andesitic tongue, but it is not andesitic east of the Big Snowy Mountains nor east and east-northeast of Red Lodge, Montana. Thom and Dobbin (1924, p. 493) state that the andesitic or volcanic material in the Lebo is still evident in the Bull Mountains and east of the Porcupine Dome. Throughout the area enclosed by the Lebo andesite line (Pl. 7) the formation is noticeably andesitic. The Lebo equivalent in the western part of the Crazy Mountain basin is the lower conglomerate sequence of the area east of the Bridger Range; it may include part or all the overlying sandstone and conglomerate units. Paleontological data are lacking.

SOURCE AREAS AND DEPOSITIONAL BASIN: Northwest and north of the Big Snowy Mountains, Upper Cretaceous and Paleocene sediments are nonandesitic, suggesting that in this area drainage was eastward and southeastward.

Similarly, the fact that the Upper Cretaceous and Paleocene sediments of the Bighorn basin in north-central Wyoming are nonandesitic indicates that their source areas did not include appreciable andesitic volcanics. Further, isolated remnants of Upper Cretaceous sediments in the Cinnabar coal field near Gardiner, Montana, and in the upper Ruby River valley northeast of Lima, Montana, are not notably volcanic, whereas equivalent beds of the Crazy Mountain basin are. This indicates drainage in south-central and much of southwestern Montana was toward the north and northeast during Late Cretaceous and Paleocene time.

In the vicinity of, and intruded by the Boulder batholith, are large, thick deposits of volcanic material. Between Jefferson Canyon and Helena, certain thick remnants of volcanic sediments, pyroclastics, and flows have been assigned a Late Cretaceous (?) age and in some instances (Alexander, 1951, Unpublished Ph.D. Thesis, Princeton University; Berry, 1943, p. 1–29) correlated with part of the Livingston formation. These deposits are most likely the remains of the voluminous volcanics which were the source of much of the Livingston sediments.

East of the Bridger Range, the conglomerates of the Livingston formation include Precambrian metamorphics, which indicates that erosion had cut deeply into near-by source areas. However, no Beltian material was identified. This, in conjunction with the location of the volcanic materials, suggests that much of southwestern Montana was a source area for the Livingston sediments but that the source consisted of two distinct parts.

One part, south of the old Beltian shore line, was positive and was the scene of limited volcanic activity such as local outbursts along major fracture zones. Erosion in this area, therefore, easily removed the minor amounts of volcanics and cut into older sediments and finally into Precambrian metamorphics.

The other part of the source area, north of the old Beltian shore line, was negative and was the site of intensive volcanic activity, especially in the vicinity of the present Boulder batholith. In this area such large quantities of volcanic materials were present that erosion never cut far into older sedimentary bedrock. Thus the Beltian sediments were never exposed to erosion.

Debris from pre-Livingston rocks, the coarse conglomerate material, appears high in the Livingston formation. According to the paleontological data, these conglomerates are Paleocene. Therefore the first strong pulse of deformation and uplift which allowed deep cutting in the source areas was Paleocene. A period of comparative quiescence allowed accumulation of 2000 feet of andesitic sandstones; renewal of compression and acceleration of erosion, probably during Paleocene time, produced the materials of the uppermost conglomerate sequence in the area east of the Bridger Range.

It is not clear whether these pulses of deformation produced any structures in the Bridger Range itself, but probably they did. Along Battle Ridge, for instance, aerial photographs suggest unconformity by apparent thinning or cutting out of beds within the sandstone be-

tween the two conglomerate units. The unconformable relationship was not proved in field work, however. Similarly, on the north end of the Chestnut anticline in the southern part of the Bridger Range area, abrupt changes in dip and strike of beds within the Livingston suggest unconformable relations, but again this was not proved. These relationships, if unconformable, indicate strong structural movements in the immediate vicinity of the Bridger Range during deposition of the Livingston.

On the other hand, much of the deformation in the area evidently postdates the youngest conglomerate of the Livingston because these beds have dips up to 45°–50° on the northwest limb of the syncline.

The Livingston formation is one key to the evolution of Laramide orogeny in southwestern Montana. The voluminous andesitic debris in the Upper Cretaceous sediments marks the approach of mountain-building activity. The periodic encroachment of terrestrial material into the depositional basin likewise marks major pulses of uplift, volcanic activity, and erosion in the source areas to the west. Thus three major pulses are recorded in the sediments: Judith River, Lennep, and Lebo. With each pulse the andesitic debris reached farther into the depositional basin. In the Lebo, Laramide orogeny may have reached its culmination along the west rim of the Crazy Mountain basin and in areas to the west, as indicated by the wide expanse of the Lebo andesitic tongue and by the conglomeratic Lebo equivalent in the area east of the Bridger Range.

Tertiary

Post-Livingston Tertiary deposits are exposed in the vicinity of Foster Creek (Pls. 1, 2). Beds assigned to the Tertiary also crop out in the southern part of the map area, north and south of East Gallatin River. Elsewhere along the west side of the Bridger Range, Tertiary beds are covered by Quaternary fan and river gravels. The limits of Tertiary outcrops shown are approximate.

The Tertiary material consists of pure, white acidic volcanic ash, buff-colored silt, and conglomerates composed mainly of Paleozoic and Belt pebbles and cobbles. It is estimated that at least several hundred feet of Tertiary is exposed in the area along Foster Creek. These beds are coextensive with Tertiary deposits near Menard, on Dry Creek, which have been described by Klemme (Unpublished Ph.D. Thesis, Princeton Univ., 1949, p. 68):

> "In the vicinity of Dry Creek near Menard, Montana, a section of Tertiary sediments 1700 feet thick (estimated), is separated from older rocks by an unconformity. Another unconformity separates the section into two units. The lower unit consists of 450 feet of interbedded fresh-water limestone and tuffaceous claystone and a buff vitric-lithic tuff. The upper unit consists of a basal sequence of conglomerate and interbedded vitric-lithic tuff overlain by a massive sequence of vitric tuff. Paleozoic and Livingston formation pebbles are the main constituents of the conglomerate. The lower unit contains fragmentary remains of Oligocene mammals, but no fossils were obtained from the upper part."

The base of the beds at Foster Creek is not exposed, and Quaternary fan deposits and alluvium unconformably overlie the Tertiary strata. No fossils have been found in these beds, but they are dated as Tertiary (probably Oligocene and/or Miocene) on the basis of lithologic similarity and continuity with other known Tertiary deposits in contiguous areas.

Tertiary strata in the southern part of the map area are well exposed in the bluffs along the north side of East Gallatin River. There they consist of coarse conglomerates with interbedded silt, sand, and ash beds. The series is estimated to be more than 400 feet thick, and the base is not exposed.

Volcanic pebbles make up the largest part of these conglomerates, and some chert and Paleozoic limestone pebbles are present. No fossils were found in these beds, but they are considered Tertiary and related to the Bozeman lake beds, which are Oligocene and Miocene (Wilmarth, 1938, p. 245–246). Peale (1896), Iddings and Weed (1894), and Skeels (1939) mapped the same exposures within the map area as Bozeman lake beds. Beds south of the East Gallatin River are also mapped as Tertiary following previous writers. The abundant volcanic pebbles in these beds indicate that the source lay to the south in the vicinity of the Gallatin Range which is capped by lavas and intruded by igneous material related to the

Yellowstone Park early basic breccias (late? Eocene).

These beds dip gently east and abut against the Livingston formation along what is considered a fault line. The contact along this line is not exposed. North of the East Gallatin River, the Tertiary beds have been beveled by a north-northwest-sloping erosion surface (Pl. 5, fig. 1). This surface passes beneath the Quaternary fan deposits along or just north of the lower part of Bridger Creek.

Quaternary

Quaternary deposits in the Bridger Range area consist of gravel fans, river gravels and muds, talus, terrace gravels, morainal and outwash deposits, and avalanche debris. The writer has mapped only the fan and river deposits on the west side of the range.

Along the west flank of the range poorly consolidated fan deposits, mainly gravels derived from Paleozoic and Precambrian rocks of the Bridger Range, lie unconformably upon probable Oligocene and/or Miocene beds. These are fan deposits, in part coalescing, which slope gently away from three centers: (1) the mouth of Middle Cottonwood canyon, (2) the Ross Creek-Potter Gulch area, (3) the mouths of North Cottonwood and Mill Creek canyons. Of these fans, (3) is the most conspicuous and is composed of much coarser bouldery debris than are (1) and (2). There is little evidence of glacial action on the headwaters of Middle Cottonwood Creek, which is the main drainage feeding fan (1). Similarly, most of the streams draining into fan (2) show little evidence of glacial action. On the other hand, there is abundant evidence of glacial action on the headwaters of the streams draining into fan (3). Thus there seems a general correlation between prominence and coarseness of individual fans and the amount of alpine glaciation on the headwaters of their feeder streams.

These fans have been incised, by present streams, to at least 100 feet in many places. Little coarse debris is carried by these streams today, even in high-water periods. The fans are therefore not being aggraded but rather are being dissected. The fans were probably built during intensive Pleistocene glaciation, probably the Wisconsin stage. These same deposits have been mapped as Bozeman lake beds (Peale, 1896; Montana State Geologic Map, 1944) and as terrace gravels (Skeels, 1939).

River gravels and muds occupy much of the bottom lands adjacent to the East Gallatin River and along the lower part of Bridger Creek. Above Bridger Canyon, on Bridger Creek, fluvial deposits have accumulated as a result of the local base level provided by the resistant Madison limestone at the canyon. Locally, also in the Bridger Creek area, stream-cut terrace surfaces are veneered by gravel. All these deposits are probably recent.

East and west of the crest of the Bridger Range, from Ross Peak north to the headwaters of Mill Creek and Frazier Creek, are evidences of alpine glaciation. Local terminal and lateral moraines and outwash deposits are preserved at the headwaters of north fork of Brackett Creek, Fairy Lake Creek, and Frazier Creek on the east side of the mountains, and at the headwaters of Mill Creek, Bill Smith Creek, North Cottonwood Creek, and Dry Fork of Ross Creek on the west side of the range. Along the east side of the mountains, from Ross Peak northward to the headwaters of Frazier Creek, a blanket of varying thickness of unsorted, extremely coarse debris covers much of the Mesozoic bedrock. This debris contains irregular blocks up to 30 feet in diameter and is heterogeneous, indicating that it is not water-worked. In places this uneven blanket is more than 250 feet thick. This coarse material appears to be a combination of morainal material, outwash, and avalanche debris. Avalanche debris connotes material moved from its source high on the flanks of the mountain for a mile or more from the foot of the mountain, by means of avalanching on the top of extensive snow fields or snow banks. This process operates on a small scale in many of the present cirques and valleys in the Bridger Range. Such a process was a major means of transportation of rock debris during Pleistocene time.

South of Ross Peak there is little evidence of glaciation; there are no well-developed cirques or glacial moraines. Locally, as on the headwaters of Middle Cottonwood Creek and Bostwick Creek, on the west side of Bridger Peak, small, poorly developed cirques are present. In spite of the absence of strong alpine glacia-

tion in the southern part of the range, coarse unsorted material forms an irregular blanket over much of the Mesozoic bedrock on the east side of the mountains. This material is considered avalanche debris deposited in the manner described for the area to the north. Probably these deposits formed during the Wisconsin stage of the Pleistocene.

Locally, large talus cones and sheets obscure the bedrock geology in the cirques and valleys of the Bridger Range.

Igneous Rocks

Sill

A differentiated composite sill extends from sec. 34, T.2 N., R.6 E., northward for 8 miles along the west side of the crest of the range, to sec. 5, T.2 N., R.6 E. Throughout its extent this intrusive is confined to Cambrian rocks. Near its southern end the sill lies within the Park shale where it is cut by a minor fault. North of the fault, the sill lies within the upper part of the Meagher formation. Beyond the boundary between Plates 1 and 2 the sill migrates downward in the section into the Wolsey shale, which position it maintains northward beyond the Cross Range fault to Mill Creek. For 10–20 feet from the intrusive the shales have been metamorphosed to hard, gray-green hornfels, and limestones are marmorized 1–2 feet from the contact.

The sill is 80–200 feet thick and locally is differentiated into a thick lower layer of biotite-augite diorite and a thinner, upper syenite layer. At the upper contact is a 1–2-foot chilled layer of biotite-augite micordiorite.

Cursory thin-section studies show some of the plagioclase of the upper chilled layer in radiating sheaflike masses, and some of augite replaced by calcite. The interstitial material of this zone consists of fine- to medium-grained plagioclase and biotite with scattered magnetite granules. Accessory minerals include orthoclase (?) and a small amount of interstitial quartz. Chloritic material and much carbonate appear as secondary replacements.

The thick basal layer of the sill is a coarse-grained biotite-augite diorite with accessory quartz, olivine, anorthoclase, and minor amounts of apatite, magnetite, and pyrite. Plagioclase is altered to white mica in some places. Greenish-brown biotite occurs as large plates and euhedral books.

The acidic layer appears to be hypidiomorphic-equigranular syenodiorite consisting predominantly of a cloudy altered plagioclase with abundant orthoclase and accessory biotite, quartz, and apatite. Minor amounts of magnetite, sphene (?), and pyrite are present.

The age of the sill cannot be determined because it intrudes Cambrian rock only. It is offset by faults and is mineralogically similar to the dike along the Pass fault. The dike cuts beds as young as Colorado, and the lithologic similarity suggests that the sill may be younger than Colorado also. Therefore a Montanan-Lancian age is suggested for the intrusive.

Dike

A locally composite dike extends along Pass fault from sec. 25, T.1 N., R.6 E., north-northwest to the east-central part of sec. 10, T.1 N., R.6 E. At its northern end the dike lies within the Amsden-Big Snowy unit; toward the south it gradually transgresses the beds until it intrudes the basal part of the Colorado formation on the south branch of Maynard Creek. Locally this intrusive is about 500 feet thick; generally it is less than 300 feet and thins toward the extremities (Pl. 1). In the NW¼ sec. 14, T.1 N., R.6 E., the dike contains a medium-grained acidic phase which is probably the result of composite intrusion. The country rock on opposite sides of the dike has been offset by fault movement which apparently preceded intrusion.

The dike is a porphyritic, panidiomorphic, olivine-augite-biotite diorite whose texture suggests that of a lamprophyre (A. F. Buddington, Personal communication). Phenocrysts consist of magnesian olivine, up to 5 mm in diameter, augite, and biotite. The groundmass consists predominantly of plagioclase and some biotite, with accessory apatite and magnetite. Olivine phenocrysts are altered to fibrous serpentine and have rims of magnetite.

The acidic portion of the dike is a biotite-augite syenite, predominantly a meshwork of orthoclase (?) and plagioclase with granules of augite, laths and plates of biotite, and accessory apatite and magnetite. The contact between the syenite and diorite is not exposed.

Contacts of the dike with the country rock are normal intrusive type. No shear zones were found within the dike, and the texture of the rock suggests no tectonic movement with or following intrusion. Therefore intrusion was probably post-faulting.

Structure

Regional Setting

The Bridger Range is about 50 miles east of the main exposed portion of the Boulder batholith and constitutes part of the Rocky Mountain front. It lies on the west and southwest rim of the Crazy Mountain basin and is subdivided into northern and southern portions by the Ross Pass fault zone. South of the fault zone there are no Belt sediments, and the area constitutes part of what Sloss and Moritz (1951, p. 2135–2169) call the Wyoming shelf. The sedimentary record indicates that the shelf area has been dominantly positive and generally stable. In contrast, north of the fault zone a thick sequence of Belt arkoses forms the southern marginal facies of the east-west Beltian embayment described by Sloss (1950, p. 430–431). During late Devonian time an east-west positive axis was active, coinciding with the present Little Belt-Big Snowy line of uplift (Andrichuk, 1951, p. 2384), and during Mississippian time an east-west trough partially overlapped the position of the Beltian embayment. Thus there has been a differentially mobile east-west belt across central Montana since late Precambrian time.

Many of the present structural features of the plains region bear a marked west-northwest alignment: the north flank of the Beartooth Range, the Nye-Bowler line, the Lake Basin fault zone, the south flank of the Big Snowy Mountains, the Cat Creek fault zone, the fault flanking the south side of the Little Belt Mountains, and the whole Little Belt-Big Snowy-Porcupine dome uplift (Structure Contour Map of the Montana Plains, Dobbin and Erdmann, 1946).

The Ross Pass fault zone in the Bridger Range lies at the juncture of three major tectonic elements in Montana: the Wyoming shelf, the Montana Plains (especially the Crazy Mountain basin), and the intensely deformed element which constitutes the area west and north of the Beltian shore line. In large part the structural complexity of the Bridger Range area can be ascribed to this location.

Compressional Features of the Bridger Range

General.—The Bridger Range is structurally unique in the tectonic complex of the Montana Rocky Mountains. It does not appear to be the east limb of a large simple anticline, as previous writers supposed, nor does it seem to constitute the draped rampart of a large block uplift, for no such uplift exists. Both long linear folds and simple low-angle thrusts, the usual earmarks of fold-mountain type of deformation, are practically nonexistent (Pls. 1, 2).

The range, in general, consists of steeply dipping Paleozoic sediments trending arcuately from Bridger Canyon on the south to Blacktail Mountain on the north. The Paleozoic rocks are flanked on the west by a narrow strip of Precambrian metamorphics in the south half and by Precambrian arkoses in the north half; on the east they are bordered by less resistant, strongly upturned Mesozoic strata. Locally, on the west, the metamorphic rocks contain infolded and infaulted remnants of Cambrian and Devonian rocks. The west side of the Precambrian terrane is terminated by steep late Cenozoic fault-line scarps.

The structure of the northern half of the range is further complicated by a series of high-angle faults trending generally northwest, on which movement has been both by strike slip and dip slip. In the southern half of the range faults having a variety of trends and dips produce complex relationships. To facilitate description of the structural features of the Bridger Range proper—that is, in the Precambrian, Paleozoic, and lower Mesozoic rocks—the range is divided into structural segments. To the west, bedrock is obscured by Tertiary and Quaternary sediments, and to the east the segments lose their identity in the later Mesozoic sediments on the rim of the Crazy Mountain basin. From north to south along the range, the major segments are: the North Bridger, Sacajawea, Truman Gulch, Schafer Creek-Middle Cottonwood, Baldy Mountain-Bridger Canyon, and Rocky Canyon. The Rocky Canyon segment lies south of the

Bridger Range proper and includes the north end of the Chestnut Mountain anticline.

North Bridger segment.—This segment includes the area bounded on the southwest by the Cross Range fault and extends north and east to the limits of the map (Pl. 2). Except close to the Cross Range fault and in the vicinity of Blacktail Mountain on the extreme north end of the Bridger Range, the strata of this segment trend approximately N. 10°W. and N. 30°W. and dip steeply northeast under the valley of Flathead Creek or are overturned toward the east and northeast. From Flathead Pass southward to the Cross Range fault the beds trend about N. 30°W., whereas north of Flathead Pass the trend is N. 05°W.

Along a narrow zone trending generally northeast in secs. 12 and 14, T.3 N., R.5 E., east of Blacktail Mountain, the beds change abruptly from strongly overturned on the southeast side to gently north-dipping on the northwest side of the zone. No displacement was noted although strong crumpling and crushing of beds is present at the juncture. The dip of the strata west of the zone steepens rapidly northward; near the margin of the map area the beds are overturned. This rapid northward overturn may be related to rupture at depth (Pl. 8, M-M′). This suggests northward movement of the North Bridger segment, which is important in the interpretation of the structural history.

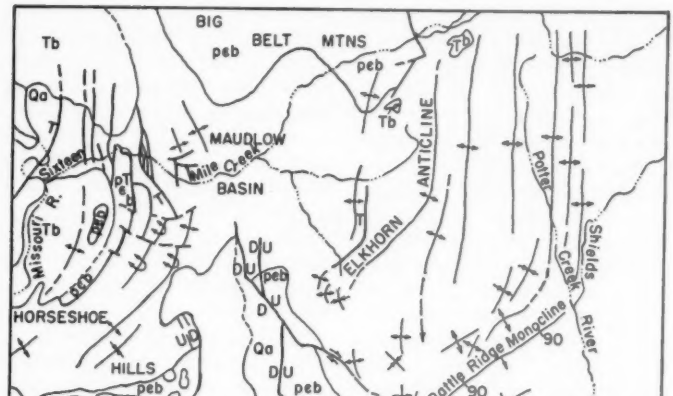
The Cross Range fault, one of the most prominent and striking structural features in the Bridger Range, can be traced from Fairy Lake Creek on the southeast to Mill Creek on the northwest, and it can be extended as far northwest as the mineralized zone in sec. 1, T.2 N., R.5 E. Farther northwest its presence is postulated on the basis of physiographic evidence. Continuous with the fault, and extending from the mineralized zone northwest to its junction with Pass Creek, is Johnson Canyon, which further coincides with the face of the mountains north of Pass Creek. The scarplike face (fault-line scarp) is linear, truncates the structures of the mountains, and marks an abrupt termination of an old erosion surface preserved on interstream ridges in the mountain area. The absence of pre-Tertiary outcrops in the area immediately southwest of the fault-line scarp further suggests that bedrock now lies considerably below the present Quaternary and Tertiary sedimentary cover.

West of Blacktail Mountain, and less than 1 mile east of the mountain front is an arcuate north-trending fault of small displacement. This fault and the fault-line scarp at the mountain front have been interpreted (Klemme, 1949, Unpublished Ph.D. Thesis, Princeton University) as high-angle underthrusts dipping steeply east, though the westernmost fault is probably part of the normal fault system terminating the west side of the Bridger Range.

The Cross Range fault can be traced southeast as far as the bed of Fairy Lake Creek. South and southeast of there outcrops are blanketed by glacial and avalanche debris and by heavy timber. In line with the fault and beyond the covered area, however, near the forks of Brackett Creek, the lower beds of a thick sequence of Livingston conglomerates show an abrupt change in attitude (Pl. 5, fig. 2). South of the forks the conglomerates strike generally north and dip steeply east; the angle of dip decreases gradually southward, whereas north of the forks the conglomerates veer sharply to a northeast strike and to strongly overturned dips (60°–80°NW.). The latter beds are on the southwest end of Battle Ridge, a prominent monocline held up by the lower conglomerate series of the Livingston formation and extending northeastward across the Crazy Mountain basin to Wilsall on the Shields River (Fig. 4).

The plane of the Cross Range fault is rarely exposed. The northeast side is upthrown along its entire length, but the movement is a combination of strike slip (northeast side moving southeast) and underthrusting of the northeast block by the southwest block. The dip of the fault plane is not apparent over much of its length; however, the pattern made by the trace and topography indicates a dip of 60°–80°NE.

Drag of beds along the fault trace locally indicates a strike-slip movement, particularly on the northeast side of the fault in the Mission Canyon limestone in central and northwest sec. 15, T.2 N., R.6 E., and in the Quadrant, Ellis, and Morrison formations on the southwest side of the fault in secs. 15 and 16, T.2 N., R.6 E. The axis of folding in both cases in-



LEGEND

- Qa — Quaternary fans & alluvium
- Tb — Tertiary basin deposits
- Paleozoic-Mesozoic undifferentiated
- peb — Pre-Cambrian Belt Series
- Pre-Cambrian Metamorphics
- Folds
- U / D — High Angle Fault, undifferentiated
- Thrust Fault, T on H.W.
- Fault, Strike-slip dominant
- Conspicuous Fractures

SCALE

10 mi.

FIGURE 4.—GEOLOGIC MAP OF REGION SURROUNDING BRIDGER RANGE, MONTANA

dicates that the northeast block moved southeast and slightly up relative to the southwest block.

In the northeast part of sec. 15, T.2 N., R.6 E., a tightly folded syncline and anticline are interpreted as drag folds resulting from a northeastward crowding and underthursting on the Cross Range fault. In sec. 9, T.2 N., R.6 E., south of a triangulation point marked 9360 (Pl. 2), minor bedding-plane slips within

the Lodgepole limestone have dragged the beds (Pl. 6, fig. 1). The drag is considered the result of underthrust movement on the Cross Range fault.

In the extreme northeast corner of the map area (Pl. 2) a faulted anticline is shown as mapped by Klemme (Unpublished Ph.D. Thesis, Princeton Univ., 1949). Klemme considers the fault a break thrust; the pattern of the structure indicates southeastward movement in which an anticlinal fold developed and later faulted parallel to its crest. The southwest end of this fold is interesting in its relation to the larger Bridger Range uplift. Instead of plunging neatly in an ellipse-shaped nose, this fold has been appressed by a synclinal cross fold, termed the South Fork sag. South of this area, along parallel 46° N., a zone of minor folds is a southwest continuation of the northeast-trending Elkhorn anticline (Fig. 4). The fold axes of this zone curve from a general northeast to a north trend as they are traced toward the mountain front. This change in trend and the cross folding (South Fork sag) are mainly the result of crowding toward the northeast by the North Bridger segment, owing to underthrusting on the Cross Range fault, subsequent to the strike-slip movement and development of the northeast-trending folds.

The structural features related to the Cross Range fault appear incompatible with one period of movement. Two sets of drag phenomena point to separate strike-slip and dip-slip movements. Generally northeast-trending folds in the area northeast of the Bridger Range suggest a relationship to the strike-slip movement on the Cross Range fault, and deflection or appression of their trends as they approach the range suggests a later northeastward crowding related to dip-slip movement on the same fault.

Regional relationships (Fig. 4) suggest that the Sacajawea segment, North Bridger segment, and the Horseshoe Hills were once part of a large northeast-plunging anticline, with a core of coarse Beltian arkoses. This eastwardly asymmetrical anticline was separated from the south end of the much larger Big Belt uplift to the north by a sag which is the present Maudlow basin.

These larger features were apparently produced by the first major phase of Paleocene compression, and then, in the closing stages of this phase, the deformation became effective eastward through the structurally low Maudlow basin area. Facilitated by strike-slip movement on the Cross Range fault, the deformation produced northeast-trending folds in an area limited on the southeast by the Battle Ridge monocline. In the development of the latter structures the Cross Range fault functioned as a tear, probably connecting at depth with a thrust which almost certainly subtends the monocline.

A second major phase of Paleocene (?) compression from the south-southwest produced an underthrust movement on the Cross Range fault, crowding the North Bridger segment north and east. Northward overturning on the north flank of Blacktail Mountain, certain drag features along the fault, appression or deflection of the northeast-trending folds, and oversteepening of the strata along the east side of the segment are all believed to be results of the underthrust movement.

Sacajawea segment.—The Sacajawea segment includes the area south and west of the Cross Range fault and extends southward roughly to Potter Gulch and Ross Pass (Pls. 1, 2). The segment lies north of the large area of Precambrian metamorphics drained by Bostwick Creek, Truman Gulch, and Jones Creek (Truman Gulch segment). In its western portion, the Sacajawea segment is a thick series of Beltian arkoses mainly eastward- to northeastward-dipping and is flanked on the east by northeastward-dipping Paleozoic and Mesozoic sediments. The southern part of the segment is marked by various trends in the post-Beltian sediments and by a pronounced zone of faulting, here called the Ross Pass fault zone, including the Ross Peak, Pass, and Dry Fork faults, as well as other minor ruptures.

Through most of their outcrop in this segment the Paleozoic and Mesozoic strata strike approximately N. 25°W. and dip 40°–65°N.E. About 1 mile north of the boundary between Ts. 1 and 2 N., the strike of the beds changes abruptly, passing through an *S* curve as the beds approach the Ross Peak fault. Dips are eastward throughout this *S* curve and remain steep. The beds of this portion of the Sacajawea segment are terminated on the south and southwest by the Ross Peak fault, which can

be traced from the headwaters of middle fork of Brackett Creek westward along the south face of Ross Peak, thence northwest across Dry Fork of Ross Creek to the divide between Dry Fork and Limestone creeks. East of Ross Peak coarse glacial and avalanche debris and heavy timber obscure the trace of the fault.

The trace south and west of Ross Peak suggests a northeast-dipping fault plane. The variation in stratigraphic displacement along the trace suggests a maximum movement south of Ross Peak; apparent displacement diminishes to the west and north. Development of the *S* curve in the strata near Ross Peak, crumpling of the Big Snowy-Amsden and Jurassic sediments at Ross Pass, and the variation in stratigraphic displacement on the fault suggest a north-south component of compression. The southeast side of this fault moved north and down relative to the northeast side, and where the trace of the fault trends east-west, south of Ross Peak, the north side has been underthrust by the south side.

A minor fault in the NW. ¼ sec. 34 and the SW. ¼ sec. 27, T. 2 N., R. 6 E., has a pattern similar to that of the Cross Range fault (Pl. 1). This fault displaces a portion of the Cambrian beds and the sill, but becomes a bedding-plane fault in the Park shale to the south where the movement is apparently dissipated. As in the Cross Range fault the northeast side appears to have moved relatively upward and southeastward.

Along the east flank of the Sacajawea segment, across the headwaters of the north fork of Brackett Creek, a wide area is mapped as Colorado-Eagle (Pl. 1). Bedrock is almost concealed between the Kootenai-Colorado and Eagle-Livingston contacts. The latter contact can be traced from the middle fork of Brackett Creek to the east-central part of sec. 25, T. 2 N., R. 6 E., on the Cache Creek drainage. The unusual width of outcrop, between the contacts mentioned, may be explained either by folding or faulting. The structure here does not appear directly related to that of the Cross Range and/or Ross Peak faults. A syncline and anticline are postulated beneath the covered interval. In the *S* curve of the older sediments in the vicinity of Ross Peak there is indication of anticlinal and synclinal axes, and the interpretation is consistent with the regional two-phase pattern of deformation.

In the large block of Beltian arkoses and conglomerates in the western part of the Sacajawea segment, the beds generally dip 50°–70° NE. E., or SE. In the extreme western part of this block, however, west dips outline an anticlinal axis. The axis is cut off on the south end by the Pass fault but appears to continue northward (Pl. 1).

The Pass fault marks the southern extent of Beltian arkoses and the northern limit of Archean (?) metamorphic rocks in the Bridger Range. The fault can be traced from Bill Smith Creek to Ross Pass and thence south-southeast across the headwaters of Maynard Creek on the east side of the range.

Near the southeast end of the fault, apparent stratigraphic displacement along the dike is less than 500 feet but increases gradually from Ross Pass westward to Dry Fork of Ross Creek, where it is at least 1500 feet. From that creek northwestward, the stratigraphic displacement increases to about 5000 feet in the SW.¼ sec. 32, T. 2 N., R. 6 E., where Flathead sandstone is in contact with the basal portion of the thick Beltian arkoses. Immediately northwest the apparent stratigraphic displacement is opposite that shown at the southeast end of the fault. This paradox can be explained by the fact that Laramide movement on this fault was opposite that of Precambrian movement which controlled deposition of the very coarse Beltian conglomerates (Fig. 5).

The plane of the Pass fault is nearly vertical through much of its length; however, on its northwest end in T. 2 N. the fault trace on topography indicates a steep dip northeast. Much of the apparent vertical movement of the north side of the fault is a result of the anticlinal fold in the Beltian arkoses, which accounts for the rapid diminishing of stratigraphic displacement toward the southeast.

The early phase of Paleocene movement was apparently less restricted toward the north; it initiated the fold in the Beltian arkoses and produced some minor strike-slip displacement on the Pass fault. Injection of the dike accompanied or followed the latter displacement. Subsequent pressure from the south-southwest in the late phase of Paleocene (?) deformation

may have accentuated the anticlinal structure by underthrusting on the northwest end of the fault.

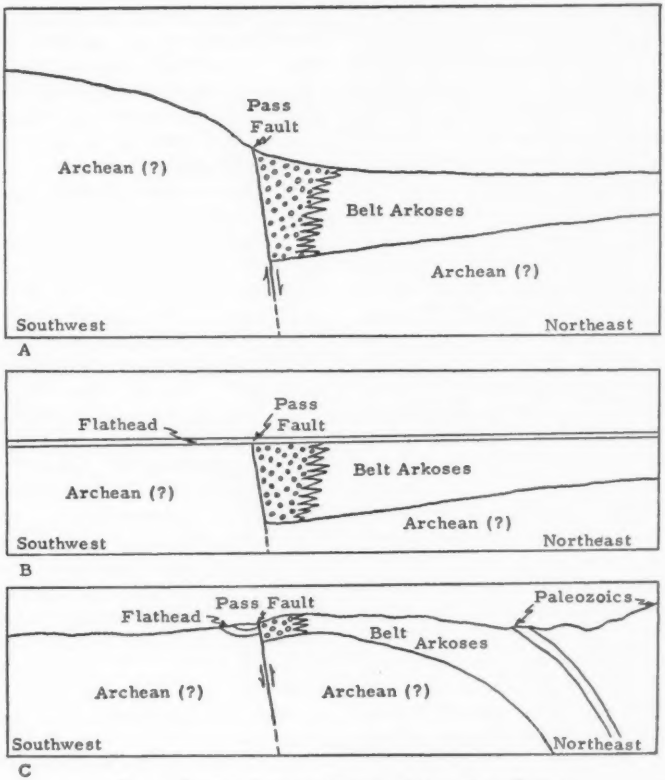


FIGURE 5.—SEQUENCE OF EVENTS ALONG PASS FAULT

Two faults of minor stratigraphic displacement bound the Sacajawea segment on the southwest. The one farther southwest, here named the Dry Fork fault, dips toward the upthrown southwest side. The maximum apparent displacement is about 400 feet. In the Middle Cambrian beds near the mouth of Dry Fork of Ross Creek are drag folds related to the thrust. Cambrian strata in the SW.¼ sec. 9, T. 1 N., R. 6 E., are overturned to the northeast, presumably also because of the late phase of Paleocene (?) movement.

Stratigraphic and structural relationships on opposite sides of the Pass fault indicate that the fault localized deposition of the Beltian arkoses. Similarly the stratigraphy suggests differential movements in the Ross Pass fault zone during Big Snowy-Amsden and Jurassic time. This zone may constitute a west-northwest extension of the Nye-Bowler zone of disturbance.

Truman Gulch segment.—The Truman Gulch segment extends southward from the Ross Pass fault zone to the Bostwick Creek fault, a major arcuate fracture in the southern part of T. 1 S., R. 6 E. It includes most of the drainage of Jones Creek, Truman Gulch, and Bostwick Creek on the west, and Maynard, Slushman, and Pine creeks on the east. The north end of the Baldy Mountain fault is the southern limit of the segment on the crest of the range.

Thus delimited, the Truman Gulch segment

is the least complicated structural unit in the Bridger Range; it is a large triangular block of Precambrian metamorphics, bounded on the east and northeast by steeply dipping Paleozoic and Mesozoic sediments. The Pass and Dry Fork faults extend into this segment. Throughout the block metamorphic foliation dips uniformly east and northeast, and near the contact it parallels the bedding of the overlying Flathead sandstone. In an anticlinal nose just north of Baldy Mountain the Cambrian and Devonian strata are strongly crumpled and shattered. East of the Baldy Mountain fault, in sec. 12, T. 1 S., R. 6 E., the upper Paleozoic and lower Mesozoic beds are tightly folded and overturned toward the east.

The Bostwick Creek fault can be traced from the north flank of Baldy Mountain, across the headwaters of Middle Cottonwood Creek, and thence westward nearly to the bottom of Bostwick Creek canyon (Pl. 1). A shear zone in the metamorphic rocks on the ridge west of Bostwick Creek probably marks the continuation of this fault. Throughout its length the trace of the fault roughly parallels the strike of the metamorphic foliation of the Truman Gulch segment. This suggests that the foliation may have controlled the orientation of the fault. Stratigraphic displacement at the southeast end of the fault is small, at most a few hundred feet, whereas on the divide between Middle Cottonwood and Bostwick creeks it is approximately 1700 feet. Allowing for present topography, the displacement farther west appears to increase somewhat. The fault plane is nearly vertical in most places and dips westward on Middle Cottonwood Creek and northward along the western portion of its trace.

The relationship of the Bostwick Creek fault to the overturning and tight folding east of Baldy Mountain suggests relative movement of the northeast side of the fault toward the southeast. Sharp curvature of the fault would preclude any extensive movement unless the plane were straighter during that phase of deformation, and were contorted by northeast-directed pressure in the later phase of Paleocene (?) deformation. In any case, the apparent displacement suggests that the Truman Gulch segment was lifted and tilted relatively eastward. Drag in the Meagher limestone adjacent to the Bostwick Creek fault in sec. 33, T. 1 N., R. 6 E., also suggests such movement.

Schafer Creek-Middle Cottonwood segment.—The Schafer Creek-Middle Cottonwood segment includes the area south of the Bostwick Creek fault and northwest of the Lyman Creek fault and Baldy Mountain fault. This segment consists of Precambrian metamorphics with infolded and infaulted remnants of Paleozoic strata (Pl. 6, fig. 2).

A faulted anticline occupies the center of the segment, and erosion of its core has extensively exposed Precambrian metamorphic rocks. The anticline is flanked on the west by a syncline containing Jefferson strata along its axial portion. On the east it is flanked by two fault-superposed synclines, and on the southeast by a high-angle reverse fault of small displacement, here called the Sypes Canyon fault, which disturbs a normal Flathead to Mississippian sequence. This sequence is terminated on the southeast by the Lyman Creek-Baldy Mountain fault. The fault along the west side of the central anticline dips steeply westward and is upthrown on the east. The apparent stratigraphic displacement along most of its length is about 500 feet and dies out southward. The fault between the two synclines in the northeast part of the segment is postulated on the basis of distribution and attitude of sedimentary units. The west limb of the syncline east of the fault is sharply overturned toward the east.

South of the syncline a gentle anticlinal flexure parallels the upper part of Middle Cottonwood Creek. Erosion has exposed a small area of Precambrian metamorphic rocks in the center of the fold. The two synclines and this anticlinal flexure die out southward.

The north-northeast-trending Sypes Canyon fault is postulated largely on the basis of shear zones and radical changes in dip and strike within the Jefferson formation. It is a high-angle reverse fault having a small upward displacement on the east side.

A west-northwest-trending fault, a continuation of the Bridger Creek fault, terminates the Schafer Creek-Middle Cottonwood segment on the southwest. Along this fault Cambrian, Devonian, and Mississippian strata abut against Precambrian metamorphic rocks.

North of the fault the Cambrian and Devonian beds are strongly crumpled and sheared, and at least two minor, northwest-dipping thrust faults cut out parts of the stratigraphic sequence. These thrusts die out rapidly to the north. The western part of the segment is composed of Precambrian metamorphic rocks which dip parallel to the bedding of the overlying Flathead sandstone. In the southern part of the segment, where the Fathead strata are vertical or overturned toward the east, the metamorphic foliation is also west dipping.

Most of the structural features of the Schafer Creek-Middle Cottonwood segment are believed to have developed during the earlier phase of Paleocene deformation. This conclusion is suggested because the folds and faults trend generally north or north-northeast and are, therefore, incompatible with an origin directly related to application of force from the south-southwest.

The westward dip of the fault along the flank of the central anticline and apparent stratigraphic displacement indicate a normal type of movement, yet this is not a normal fault in that tension was at right angles to the strike of the fault. The relation of this fault to the Bostwick Creek fault indicates an origin prior to or contemporaneous with the Bostwick Creek fault. Drag on the sharply curved portion of the Bostwick Creek fault may have carried a fault-bounded sliver of the Schafer Creek-Middle Cottonwood segment upward as the Truman Gulch segment rose and tilted eastward. This would explain both the apparent normal fault along the west side of the central anticline and the thrust fault separating the synclines to the east. Such a theory also suggests that the lift on the Truman Gulch segment was applied along its western margin rather than along the Bostwick Creek fault and that the structures developed during the earlier phase of Paleocene deformation. During the same phase the Sypes Canyon fault developed as a backthrust.

Baldy Mountain-Bridger Canyon segment.—The Baldy Mountain-Bridger Canyon segment includes the area southeast and south of the Baldy Mountain-Lyman Creek fault and extends as far as the low hill of Paleozoic and lower Mesozoic rocks south of Bridger Canyon. A small wedge of Precambrian metamorphic rocks, northwest of the mouth of Lyman Creek, is included in this segment. The main body of the segment consists of Paleozoic and Mesozoic sediments overturned toward the southeast except in the northernmost part.

The Lyman Creek-Baldy Mountain fault is the most perplexing one found in the Bridger Range. It can be traced northeastward along Lyman Creek and thence northward along the west side of the crest of the range, to the northwest flank of Baldy Mountain. A fault on the east side of Baldy Mountain, which may be connected with the Lyman Creek-Baldy Mountain fault, can be traced from the headwaters of Beasley Creek northward to the crest of the range. Heavy talus from the Lodgepole limestone obscures the critical bedrock relations on the north flank of Baldy Mountain. At its southwest end, the Lyman Creek fault is cut by the fault which terminates the Schafer Creek-Middle Cottonwood segment. Near the juncture of these two faults a large spring furnishes part of the water supply for the town of Bozeman.

Lower Paleozoic strata southeast of Lyman Creek are overturned toward the southeast. This overturning is most marked at the sharp bend in the fault trace, near the head of the creek. The overturned Sappington and lower Lodgepole beds here dip about 30°NW. but resume less overturned dips immediately to the southeast. North of the bend, the strata east of the fault are vertical and gradually assume normal east dips of decreasing magnitude as they are traced north, until on the summit of Baldy Mountain the Lodgepole limestones are nearly flat lying.

Northwest of Lyman Creek, strata from Devonian to Jurassic (Ellis) dip steeply southeast. A small wedge of strongly sheared Mission Canyon limestone lies between the Ellis beds and the Sappington formation at the sharp bend in the fault trace. West of the fault Mississippian strata dip steeply east as far north as the southwest corner of sec. 11, T. 1 S., R. 6 E. Northward the eastward dip of those beds diminishes to zero, and beds dip up to 15°W. on the northwest flank of Baldy Mountain.

Along Lyman Creek the dip of the fault was not ascertained. The trace north of its sharp bend on Lyman Creek suggests that part of the fault dips 20°–40°E. The trace of the fault

on the east side of Baldy Mountain suggests a southwest dip, possibly as low as 10°.

The backthrust interpretation (Pl. 8, A-A′, B-B′, C-C′) of the Lyman Creek-Baldy Mountain fault is the more logical of two possible explanations of the stratigraphy and structure. The other possibility requires overthrust displacement of 11,000–15,000 feet on the fault along Lyman Creek, strike-slip movement along the north-south portion, and underthrusting on the north end. This alternative theory also requires a drastic rotation of the Schafer Creek-Middle Cottonwood segment. Evidence of such rotation should have been left in the form of transcurrent strike-slip faults and shear zones; no such structures were found. The backthrust interpretation is favored because it is consistent with the features along the fault, except possibly the strong overturning of Sappington and lower Lodgepole strata at the sharp bend, near the head of Lyman Creek. This overturning is drag along a backthrust which curves back on itself toward the surface. At depth the roots of the backthrust merge with the base of the Cambrian. The sequence of events is: (1) development of a large overturned anticlinal fold by the early phase of Paleocene compression; (2) movement of the Paleozoic and younger sediments away from or out of the downwarped basin by slip along the Precambrian and Paleozoic contact, with ultimate shearing across the overturned anticlinal structure; and (3) impinging of a thrust mass on the south end of the range, near Bridger Canyon, during the later phase of Paleocene (?) deformation. This thrust mass pushed the Baldy Mountain-Bridger Canyon segment northward, accentuating the backthrust relationship in the vicinity of Baldy Mountain, and producing the fault on the east side of that mountain. The northward movement of the segment in the final stage was facilitated by strike slip on the southern part of the Lyman Creek-Baldy Mountain fault, by bedding-plane slip at the Mission Canyon-Amsden contact on the headwaters of Beasley Creek, and by minor crumpling or drag folding in the Kootenai, Colorado-Eagle, and Livingston beds between Beasley and Place creeks.

Toward the south end of the segment, near Bridger Canyon, the Paleozoic-Mesozoic beds become strongly overturned, locally to complete recumbancy, and are so badly sheared by localized thrusting that it was impossible to show all the faults at the scale used. Several faults which cut obliquely across the overturned and thrust strata are shown on Plate 1. These offsets are due to the west wall of each having moved down relative to the east wall. Cross sections F-F′ and G-G′ (Pl. 8) illustrate the surface relations in the area and the subsurface interpretations made therefrom. The localized thrusting dies out quickly to the north away from the Bridger Creek-Bear Canyon fault, as do the minor thrusts at Sypes Canyon. Intensity of thrusting and overturning is greatest south of Bridger Canyon. These features are local drag effects of a large overriding mass which moved northeastward during the later phase of Paleocene (?) deformation. A remnant of this mass is represented by the small wedge of Precambrian northwest of the mouth of Lyman Creek. This wedge is terminated on the south and west by normal fault escarpments and contains a small infolded and infaulted synclinal remnant of badly sheared Middle Cambrian strata.

Rocky Canyon segment.—The Rocky Canyon segment includes the area south of the Bridger Range proper and east of the Bridger Creek-Bear Canyon fault and extends southward to the limits of the map area. The rocks are largely Upper Cretaceous sediments, except in the core of the Chestnut Mountain anticline, which is shown as mapped by Skeels (1939) with slight modification.

Certain features establish the presence of the Bridger Creek-Bear Canyon fault: (1) On the north end a scarplike face cuts obliquely across Precambrian and Paleozoic bedding. South of Bridger Canyon topographic saddles, gullies, and changes in slope mark the contact of Tertiary beds on the west with Cretaceous beds on the east. (2) Intense shearing, oxidation, and calcitization occur at or near the east side of the same contact. (3) Strata are strongly overturned and crumpled as much as three quarters of a mile east of the contact.

The deformation of the Paleozoic and lower Mesozoic strata at Bridger Canyon has been described; southward the Colorado-Eagle unit has been strongly overturned so that it dips 20°–30°W. and NW. North-northeast beyond Bridger Creek, however, these same beds

quickly resume steep-overturned to normal dips. Southward along the fault the Livingston formation is tightly folded and sheared in a zone up to three quarters of a mile wide, east of the Tertiary-Livingston contact. East of the strongly deformed zone the Livingston beds quickly flatten and show only gentle flexures.

A projection of the Bridger Creek-Bear Canyon fault to the south-southeast, across the East Gallatin River, coincides with the easternmost of two thrust faults mapped by Skeels (1939) in the area east of Bear Canyon Creek. The north end of the Chestnut Mountain anticline is included in the present map to show the connection between the Bridger Range area and the Trail Creek-Canyon Mountain area of Skeels. That fold trends north-northwest, and Mission Canyon limestone is exposed along the axial portion. On the north wall of Rocky Canyon, which cuts at right angles across the fold, the upper Paleozoic and Mesozoic strata dip gently northward. To the north, toward the plunging end of the fold, the dip steepens until along the Eagle-Livingston contact beds are nearly vertical and in some places overturned. This northward overturning is significant in that it supports the theory of north-northeast movement in the later phase of deformation in this area.

South of Rocky Canyon the west side of the Chestnut Mountain anticline is tightly squeezed, possibly owing to crowding by the thrust block which impinged on that side. Northward swing of the Eagle-Livingston contact in the extreme southeast corner of sec. 14, T. 2 S., R. 6 E., suggests drag along the Bridger Creek-Bear Canyon fault. This drag and the overturning of the north and northeast sides of the Chestnut Mountain anticline suggest a general northeast movement of the overriding block. Areally restricted intensive shearing and minor thrusting are related to the same movement. Apparently the strata in this area had been folded by the earlier phase of deformation and dipped so steeply that a small amount of drag produced by the later overthrusting on the Bridger Creek-Bear Canyon fault completed the overturning and produced other localized deformation.

Structural History

(1) The Bridger Range is at the juncture of three major tectonic elements along the south margin of an east-west mobile zone.

(2) The Pass fault was active during Proterozoic time and controlled deposition of the coarse Beltian arkoses, and was active again during Laramide deformation. The fault may be the west-northwest continuation of the Nye-Bowler fault zone.

(3) Tectonic activity related to the Laramide orogeny began in Late Cretaceous time and became increasingly severe in a series of pulses probably culminating in late Paleocene time with major folding and faulting in the Bridger Range and adjacent areas.

(4) An early Paleocene pulse of deformation caused uplift in the area west of the Bridger Range, erosion of which produced the coarse materials of the lower conglomerate sequence in the Livingston formation.

(5) A period of relative quiescence ensued during which the sands and sporadic gravels of the upper sandstone unit of the Livingston were deposited.

(6) As the latter materials were being laid down the first of the pulses of deformation which folded and faulted the rocks of the Bridger Range area began to lift Battle Ridge into its monoclinal form. The renewed uplift in the areas to the west, and accelerated erosion produced the coarse conglomeratic material of the uppermost Livingston beds. Uplift of the Battle Ridge monocline continued during deposition of these Livingston beds and produced the present monclinal structure as well as certain other major structural features within the Bridger Range itself.

(7) Structural features in and near the map area are incompatible with compression from one direction and suggest two distinct pulses from different directions. The first of these is mentioned under (6), and the features related to it can be attributed to a general east-west compressive force or to components thereof, as explained in the description of individual features. This pulse was almost certainly of Paleocene age. The second pulse is due to general south-southwest–north-northeast compression. How much later this pulse was than the preceding one and whether or not it occurred

during Paleocene time is not known. It may have been contemporaneous with northeastward movements recorded along the northeast side of the Beartooth Mountains and with large-scale thrusting in southwestern Montana which apparently occurred in latest Paleocene or early Eocene time.

Physiography

General

The writer believes a post-deformational erosion surface developed in southwestern Montana, but there is little convincing evidence of an Eocene peneplane. Early in Oligocene time this region may have been one of gently sloping lowlands, separated by ranges similar to those existing today, through which unobstructed streams flowed, carrying their debris to far-distant depositional sites. However, beginning in Oligocene time and continuing to recent times, normal faulting and other structural movements fashioned many of the present basin areas and provided temporarily closed basins in which various portions of the Tertiary deposits were laid down. These structural movements were sporadic, and one pause, in middle Pliocene time, allowed development of the widespread Flaxville pedimentary surface. According to this interpretation of Cenozoic history drainage has generally been from west to east. Scarplike faces along the west side of the Bridger Range are interpreted as fault-line scarps, as a result of normal faulting which lifted a middle Pliocene erosion surface in the range high above its complementary portion in the Gallatin Valley.

Tertiary Basins and Post-Compressional History

Most authors agree that several sets of erosional surfaces developed in western Montana during Cenozoic time and that there has been major normal faulting during the same time interval. There is, however, much disagreement both as to the exact time of development, nature, and extent of the erosion surfaces and as to the precise time, nature, and extent of the faulting.

Some geologists believe that a so-called Eocene peneplane was developed in the Rocky Mountain region. Certainly an erosion surface was produced following Paleocene deformation, a surface which in the region of the Idaho batholith was of low relief. Accordance of summits in individual mountain ranges has been considered evidence of the former existence of the peneplane; the variation in average elevation of the different ranges is attributed to post-Eocene structural movements. The fact that most of these ranges are structural entities and that they probably had accordant summit elevations at the close of Paleocene deformation suggests that the present accordant summits are related to inherent characteristics of each range and to subsequent weathering processes and not to Eocene peneplanation. The presence of Eocene basin deposits (Douglass, 1903, p. 145–149) near Lima, Montana, and coarse alpine glacial gravels (Scott, 1938, p. 628–636) of probable Eocene age in the Gravelly Range of southwestern Montana suggests more physiographic relief than is generally attributed to peneplaned areas.

Pardee (1925, p. 38–43) believes that in early Oligocene time western Montana was a region of gently sloping lowlands separated by ranges, and that while these lowlands and mountains were being carved the region had relatively free unobstructed drainage, for no deposits of Eocene age are known in southwestern Montana, other than those mentioned. He says:

> "In Oligocene time, however, conditions changed so greatly that sediments began to accumulate on the hitherto cleanly swept lowlands. The deposition may have been the result of a change from a moist to an arid or desert-like climate . . . , or the deposits may have been caused by obstructions to the drainage, either lava flows or barriers elevated by structural movements."

The last of these possibilities is most plausible and is supported by field evidence. Pardee (1950, p. 359–406) also believes tectonic activity initiated Tertiary deposition.

On the basis of a supposed dendritic pattern to the major Tertiary sedimentary basins, Atwood (1916, p. 698–738) and others have assumed that the continental divide lay along the Big Belt-Bridger Range line of uplift in early Tertiary time, far east of its present position, and that drainage from all southwestern Montana was southward into the Snake River. Evidence presented by Tanner (Unpublished Ph.D. Thesis, Princeton Univ.,

1949) and Klemme (Unpublished Ph.D. Thesis, Princeton Univ., 1949) concerning the Big Belt Mountains and Maudlow basin area indicates that streams and, probably, the ancestral Missouri flowed east across the present line of uplift in Oligocene-Miocene time. Further, it is difficult to imagine a dendritic pattern for these western Montana basins. In general, one or more margins of the larger basins are marked by known or suspected faults. At least in part individual basins parallel major Laramide compressional trends, which have in many cases been reactivated by later Tertiary and even Recent faulting. The overall pattern is and was more nearly one of structural control than of dendritic drainage. It should be remembered that during Late Cretaceous and Paleocene time drainage was apparently from western Montana into the Crazy Mountain basin and thence eastward. If drainage was to the south into Idaho during Oligocene-Miocene time, then the earlier reversal of the direction of flow must be explained.

Regardless of direction of drainage in Oligocene time it is not necessary to suppose that deposition started in the Gallatin Valley area, west of the Bridger Range, and in the Townsend Valley area, west of the Big Belt Mountains, because of climatic changes. Structural barriers raised across the drainage would have been more effective.

The Oligocene-Miocene deposits of the Gallatin Valley-Townsend Valley basins are dominantly fine-grained silts, sands, and clays composed largely of reworked volcanic ash. Some pure ash beds are present. Toward the west front of the Bridger Range and Big Belt Mountains at least the upper Miocene materials become noticeably coarser-grained, locally consisting of conglomerates derived from the immediately adjacent mountain area. Although not exposed the Oligocene deposits may likewise coarsen near the mountain fronts.

Upward movement on the east side of a zone of faulting which extends along the west side of the Bridger Range and Big Belt Mountains would adequately explain initiation of deposition in the Gallatin and Townsend valleys and the coarsening of the Tertiary sediments toward the mountain fronts. Continued, though sporadic, faulting along this zone would determine depth of accumulation and to some extent type of sediments, as well as provide adequate source areas for the conglomeratic materials.

The individual Tertiary basins are now separated by barriers formed of pre-Eocene rocks whose heights above basin bottoms are certainly the result of post-Eocene structural movements. It seems logical that the start of deposition in each individual basin may have been the direct result of initial upward movement of these barriers. If this be the case, some of the deep canyons between basins may have resulted in part from antecedance rather than entirely from superposition as suggested by previous writers.

In the Bridger Range area the truncated Tertiary beds near Foster Creek pass southward and eastward beneath Pleistocene fans and Recent river gravels; south of Bridger Creek the beveled surface of the Tertiary strata slopes north-northeast beneath Pleistocene fans. Tertiary basin deposits underlie the Pleistocene and Recent materials all along the west side of the range. Seismic work and borehole drilling by the U. S. Geological Survey indicate that bedrock lies at variable depths but is at least 1000 feet below ground level throughout much of the valley area (F. A. Swenson and O. M. Hackett, Personal communication, 1952). This bedrock surface probably consists of Precambrian metamorphics and infolded and infaulted Paleozoic sediments. The greatest depths to bedrock in the valley most probably lie near the west front of the Bridger Range.

The scarplike faces along the west front of the range are interpreted as a product of fairly recent relative uplift of the mountain flank along a series of interconnecting high-angle faults. This interpretation is based upon: (1) the abrupt termination of a high-level erosion surface by steeper escarpment faces, which are locally aligned and have a faceted appearance (Pl. 6, fig. 3); (2) the faces which cut obliquely across the internal structures of the range in most cases and mark an abrupt change in slope from the Pleistocene fans to the Precambrian rocks of the mountains; (3) the absence of pre-Tertiary outcrops west of the range front, which indicates that older bedrock lies at considerable depth below the Tertiary, Pleistocene, and Recent materials of the

western part of the map area; (4) the seismic and drilling evidence indicating depth of bedrock; and (5) morphological similarity to and alignment with the west side of the Big Belt Mountains, along which normal faults disturb the flanking Tertiary strata.

Pliocene Erosion Surface

In the western part of the Bridger Range an old erosion surface is preserved on the interstream ridges. It is irregularly developed and ranges in relief from 200 to 600 feet. The surface showed least relief in the Truman Gulch segment and most relief in the Sacajawea segment of the range, the variation being due primarily to the degree of uniformity of rock type. The surface terminates abruptly on the east against the crestal ridge of the range. This termination varies in elevation; it increases to a maximum in the Sacajawea segment and diminishes to the north and south.

At the north end of the Bridger Range, the mountainward extension of the Blacktail bench (Klemme, Unpublished Ph.D. Thesis, Princeton Univ., 1949) terminates against the north side of Blacktail Mountain at 6400–6500 feet elevation. The Blacktail bench is the highest and oldest recognizable erosion surface in the area north and northeast of the Bridger Range and is part of what Klemme calls bench 1. He correlates this bench with bench 1 of Pardee (1925) in the Townsend Valley area. Both writers consider bench 1 equivalent to the Flaxville surface of the Montana plains. The Flaxville and equivalent surfaces are variously dated as Latest Miocene or early Pliocene (Collier and Thom, 1918; Alden, 1932); probably early Pliocene or possibly early Pleistocene (W. T. Thom, Jr., Personal communication, 1952); early Pliocene (Klemme, Unpublished Ph.D. Thesis, Princeton Univ., 1949); early Pliocene? (Atwood, 1916); middle to late Pliocene (Pardee, 1925); and late Pliocene (Mackin, 1937; Horberg, 1940). The consensus favors a Pliocene age, but there is considerable question as to whether early, middle, or late.

Along the east side of the Bridger Range a high-level erosion surface, which can be correlated with the Blacktail bench, terminates against the crestal ridge at elevations from 6400 feet at the north to 7000 feet near Fairy Lake, to 6000 feet at the south end of the range. This surface is rarely well preserved because of heavy Pleistocene outwash and avalanche debris, and destruction in large part by Pleistocene and Recent erosion. A corresponding surface bevels the Livingston formation and Tertiary deposits south of Bridger Creek, and part of it slopes north-northwest under the Pleistocene fans west of Bridger Canyon (Pl. 5, fig. 1). It appears that the erosion surface which surrounds the Bridger Range on the north, east, and south would be equivalent to the Flaxville surface and presumably is Pliocene.

The high-level erosion surface in the western part of the range corresponds with the Flaxville surface on the east side of the range in that it is the highest and the only surface found, though it may be an uplifted pre-Oligocene surface. The latter hardly seems likely, however, since a period of stability long enough to develop the widespread Flaxville surface should certainly have produced recognizable benches or levels of erosion below an uplifted pre-Oligocene surface. Since no such benches or levels of erosion are found in the western part of the Bridger Range it is assumed that the high-level surface there is correlative with the Flaxville.

The youngest beds beveled by correlatives of the Flaxville surface are upper Miocene. Morainal material, probably of Wisconsin age, lies on the surface, and the surface has been eroded and locally destroyed by glacial action. The age of the surface therefore can be early Pliocene to Pleistocene. Because development of a widespread erosional surface such as the Flaxville requires a considerable interval of stability a middle Pliocene to Pleistocene age appears favorable.

The late Pleistocene fans along the west side of the range show no signs of disturbance by movement on the normal faults. However, recent activity on faults in the same zone, north of Sixteen Mile Creek (Pardee, 1927, p. 7–23), indicate the possibility of such movements in the future.

Post-compressional isostatic uparching of the Rocky Mountain region produced block faulting by differential uplift. These movements began as early as Oligocene time and have continued sporadically up to the present, with a

major pause in Pliocene time during which the Flaxville surface was developed.

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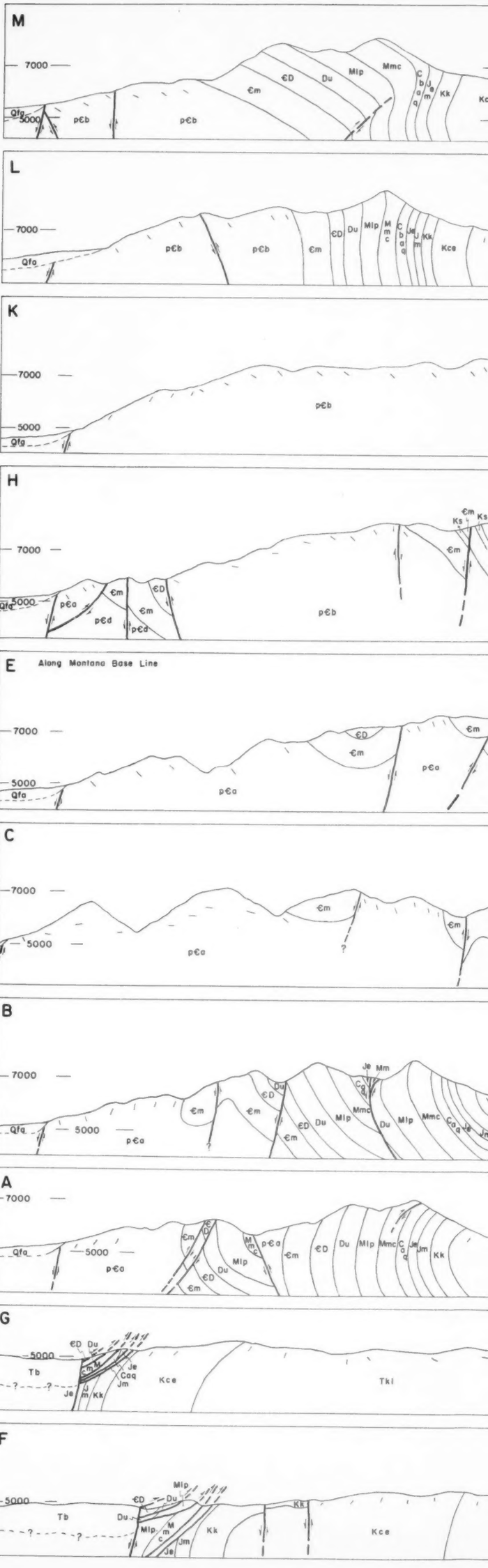
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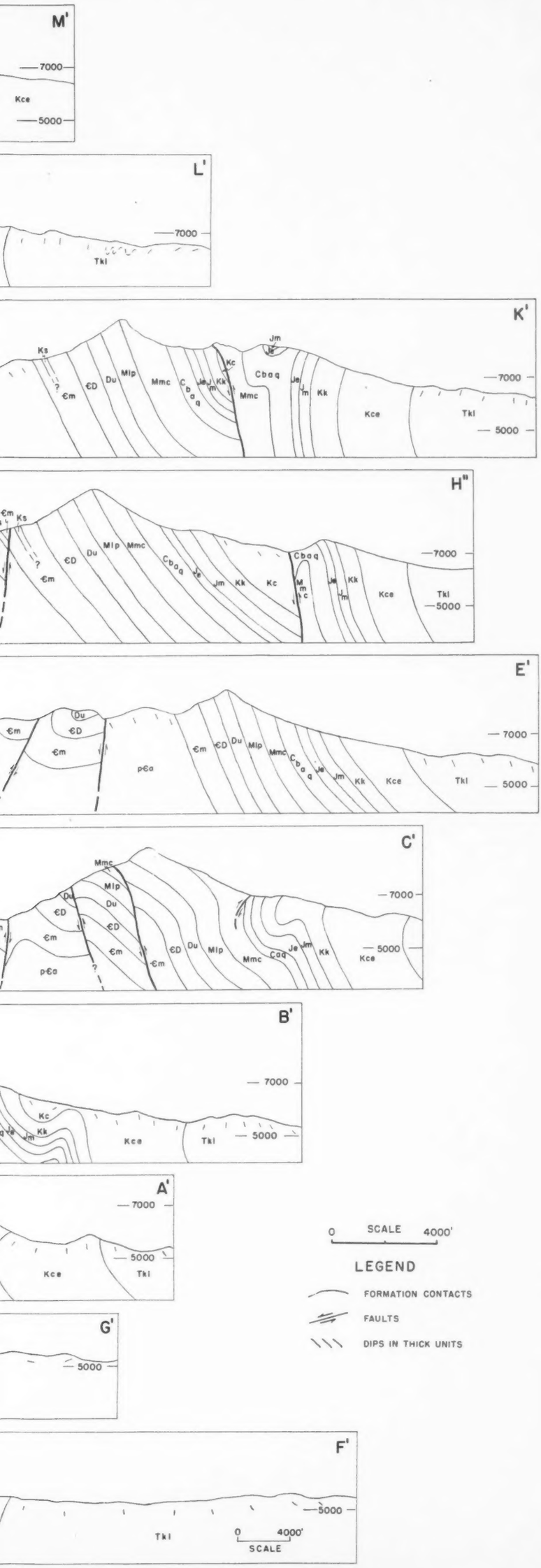

TIONS OF BRIDGER RANGE, MONTANA

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SEQUENCE OF ALLUVIATION ALONG THE LOUP RIVERS, VALLEY COUNTY AREA, NEBRASKA

BY ROBERT D. MILLER AND GLENN R. SCOTT

ABSTRACT

Alluviation along the North and Middle Loup rivers in Valley County, Nebraska, produced a series of alluvial silt beds on which are developed five interstadial soils of Brady (?), Cary and Mankato (?), Mankato (?), early Recent and late Recent age. No deposits of Illinoian age were found and in at least one place the Brady (?) soil is superposed on a gleyed horizon developed on the Sappa formation, which overlies the Grand Island formation. Post-Yarmouth erosion removed the Sappa formation and part of the Grand Island formation at most places. The characteristic Sangamon soil was not developed on uneroded remnants of the soil of Yarmouth age because of a wet, poorly drained environment. A large collection of molluscs obtained from this gley soil supports stratigraphic evidence indicating a Yarmouth age.

Alluviation during the glacial substages of Wisconsin time and soil formation during the interglacial substages were uninterrupted by erosion until after development of the soil of Mankato age. The Wisconsin deposits were dissected to below the Brady soil in places, and the resulting gullies were filled with later alluvium. The late Recent fill terrace lies 15 feet below the top of the late Wisconsin terrace. Subsequent cutting and filling in late Recent time produced another terrace 11 feet below the older terrace. The modern gully was cut during the last century.

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INTRODUCTION

This paper describes the Quaternary alluvial deposits of the Loup Rivers area, which includes parts of Valley and Howard counties, Nebraska along the North and Middle Loup rivers, and tentatively correlates these deposits with the stages of continental glaciation. In the summer of 1950 Miller mapped the geology of Valley County (in preparation) and George K. Biemesderfer mapped the geology of Howard County (Biemesderfer, in prepara-

tion). Miller and Scott revisited the Loup River area in November 1953 and made a special study of the alluvium.

Geography

Valley County is near the center of Nebraska north of the Platte River and north of Grand Island (Fig. 1). The North Loup and Middle Loup rivers, which originate in the Sand Hills, flow southeastward across the area. The boundary between the Sand Hills and the Loess Hills extends across the northeast corner of the county. The topography is characterized by deeply dissected, steeply rolling hills in the Loess Hills area and gently rolling stabilized sand dunes in the Sand Hills area. The large streams flow in wide flat valleys with broad flood plains, and the smaller streams flow in narrow alluvium-filled valleys with deep recently cut gullies.

Figure 1.—Location Map Showing Area of Terrace-Alluviation Studies

Criteria Used for Correlating Alluvial Fills

The types of evidence used to correlate alluvial fills are essentially those listed by Frye and Leonard (1952, p. 42):

"These types are: (1) lines of evidence establishing physical continuity from place to place, such as the tracing of a till sheet, a loess sheet, or a terrace surface; (2) evidence of synchronous deposition, usable only by the recognition of separated deposits made by the same volcanic ash fall; (3) evidence from the matching of similar assemblages of fossil organisms; (4) the matching or tracing of weathering effects on a former surface; and (5) the matching of similar stratigraphic or physiographic sequences."

Each of the criteria except (2) was of value in the study. Physical continuity of loess sheets and terrace surfaces established the broader relationships of the deposits but was of negligible value in tracing single alluvial fills. The lack of volcanic ash made proof of synchronous deposition impossible. Except in the Coopers Canyon gley soil, fossils were scarce and not diagnostic, but were good indicators of environment. The matching of ancient weathering profiles on buried surfaces was the primary basis for correlation from exposure to exposure. A stratigraphic column was built up and sequences of deposits and weathering profiles were then matched to extend the correlation throughout the area. Terrace surfaces are prominent and can be traced throughout the area.

Joffe (1949, p. 41) defines a soil as:

"... a natural body of mineral and organic constituents, differentiated into horizons, of variable depth, which differs from the material below in morphology, physical makeup, chemical properties and composition, and biological characteristics."

The soils in the Loup River area are northern chernozems according to Marbut (1935, Pl. 2, p. 14). Chernozem soils are defined (U. S. Dept. of Agriculture, 1938, p. 1164) as

"A zonal group of soils having a deep, dark-colored to nearly black surface horizon, rich in organic matter, which grades below into lighter colored soil and finally into a layer of lime accumulation; developed under tall and mixed grasses in a temperate to cool subhumid climate."

According to Thorp and Smith (1949, p. 121):

"Zonal soils, in general, are characterized by well-differentiated horizons and by profiles that differ markedly according to the climatic-ecological zone within which they occur."

TABLE 1.—PROPOSED CORRELATION BETWEEN TERRACE AND UPLAND SECTIONS IN VALLEY COUNTY AND KNOWN SEQUENCES OF SOILS IN THE GREAT PLAINS

Series	1 Geologic time		2 Howard and Valley Counties, Nebr. This paper (Valleys)	3 Valley County, Nebr. (Uplands)	4 Schultz, Lueninghoener, and Frankforter, 1951**
Recent			Erosion No soil Recent fill 2	Reworking of eolian sand	Soil Z' Silt and loess
			Erosion Arcadia Soil	Soil fm.	Soil Z
			Recent fill 1 Silty alluvium		Silt and loess (Cochrane "Misc. VI")
			Erosion Turtle Creek Soil* Silty alluvium	Recent loess Soil fm.	Soil Y
Pleistocene	Wisconsin glacial stage	Mankato substage	Elba Soil Silty alluvium	Eolian sand	Bignell: T-2A fill silt and loess
		Interglacial between Cary and Mankato	Ord Soil	Bignell loess	Bignell: Soil YY
		Cary substage	Silty alluvium	Wind erosion	Bignell: T-2B fill silt and loess
		Interglacial between Tazewell and Cary	Brady(?) soil	Brady soil	Brady: Soil X
		Tazewell substage	Silty alluvium	Eolian sand	Silt and loess
		Interglacial between Iowan and Tazewell	Erosion	Peorian loess Soil formation	Peorian: Soil W
		Iowan substage		Alluvial sand	Peorian: Silt and loess Todd Valley formation
	Sangamon interglacial stage		Coopers Canyon gley soil	Sangamon soil	Sangamon soil
	Illinoian glacial stage			Loveland loess Crete formation	Loveland formation (silt and loess) Crete formation
	Yarmouth interglacial stage			Soil of Yarmouth age	Yarmouth soil
	Kansan glacial stage		Sappa formation Grand Island formation	Sappa formation Pearlette ash member Grand Island formation	Sappa formation Pearlette ash, silt, sand, etc. Grand Island formation sand and gravel Red Cloud formation sand and gravel
	Aftonian interglacial stage				Fullerton formation silt and sand
	Nebraskan glacial stage				Holdrege formation sand and gravel

*Formed during climatic optimum. **Nebraska State Museum Bull., v. 3, no. 6, 1951.

The horizons that are recognized by the U. S. Department of Agriculture and which are used in this paper are the A, B, C, and G horizons, which are defined in the Soil Survey Manual (U. S. Dept. of Agriculture, 1951, p. 178–180):

"The *A horizon* is a master horizon consisting of (1) one or more surface mineral horizons of maximum organic accumulation; or (2) surface or subsurface horizons that are lighter in color than the underlying horizon and which have lost clay minerals, iron, and aluminum with resultant concentration of the more resistant minerals; or (3) horizons belonging to both of these categories."

"The A horizon, and especially the A_1, is the horizon of maximum biological activity and is subject to the most direct influences of climate, plants, animals, and other forces in the environment. In a sense, the A protects the rest of the soil; and in it many of the most important soil-building forces have their origin."

"The *B horizon* is a master horizon of altered material characterized by (1) an accumulation of clay, iron, or aluminum, with accessory organic material; or (2) more or less blocky or prismatic structure together with other characteristics, such as stronger colors, unlike those of the A or the underlying horizons of nearly unchanged material; or (3) characteristics of both these categories. Commonly, the lower limit of the B horizon corresponds with the lower limit of the solum [The A and B horizons]."

"The *C horizon* is a layer of unconsolidated material, relatively little affected by the influence of organisms and presumed to be similar in chemical, physical, and mineralogical composition to the material from which at least a portion of the overlying solum has developed."

The G horizon:

"... is a layer of intense reduction, characterized by the presence of ferrous iron and neutral gray

colors that commonly change to brown upon exposure to the air. It is a characteristic horizon in soils developed wholly or partly by gleying. This process involves saturation of the soil with water for long periods in the presence of organic matter. One may speak appropriately of a 'gley (glā) soil' but hardly so of a 'gley horizon,' since the genesis of the whole profile is involved."

At each exposure of alluvium every soil horizon was described with special attention to thickness, boundary, color, mottling, texture, structure, consistence, reaction, and special features such as concretions and fossils. The procedure was that described in the Soil Survey Manual (U. S. Dept. of Agriculture, 1951, p. 123–244).

Stratigraphy of the Valley County Area

The oldest formation in the area is the Pliocene Ogallala formation. Sand and gravel of the Grand Island formation of Kansan age unconformably overlie the Ogallala formation along the valleys. Silt of the Sappa formation and ash of the Pearlette ash member, both of late Kansan age, conformably overlie the Grand Island formation. In places a soil of Yarmouth age is developed on the Sappa. In the valleys a gley soil is developed on the Sappa formation. Because of its good development in Coopers Canyon in sec. 14, T. 15 N., R. 11 W., Howard County, Nebraska, it is called the Coopers Canyon gley soil. Sand of the Crete formation of early Illinoian age may unconformably overlie any of the earlier formations, but normally overlies the Sappa formation and is overlain by the Illinoian Loveland loess which forms the core of most hills in the uplands. The Sangamon soil is developed on the Loveland loess (Table 1). Alluvial sand of Illinoian or Iowan(?) age overlies the Loveland loess along ancient valleys in the upland. The pre-Wisconsin deposits crop out along the stream courses, in erosion scars in the uplands, and in road cuts.

The Peorian loess of Iowan and Tazewell age blankets the hills in the upland, and its valley equivalent is the lower silt bed in the Wisconsin alluvial fill. The Brady(?) soil is developed on this bed. The silt beds are not eolian alone but are mostly alluvial or colluvial. The Bignell loess of Cary and Mankato ages overlies the Peorian loess and the Brady soil in the uplands. The valley equivalent of the Bignell loess is two distinct beds of silty alluvium. On the lower bed is developed the Ord soil, named for an exposure in a nearly vertical embankment in the SW¼ NW¼ sec. 21, T. 19 N., R. 14 W. on the west side of the golf course at Ord, Nebraska (sec. 3). On the upper bed is developed the Elba soil, named for an exposure 50 yards east of the bridge on State Highway 11, sec. 14, T. 15 N., R. 11 W., Coopers Canyon, Howard County, Nebraska (sec. 7). In the northeast corner of Valley County, eolian sand of early Recent age overlies the Bignell loess. Early Recent is considered to include deposits that were laid down during the time from Mankato to the climatic optimum. Silty alluvium of early Recent age overlies the Elba soil in the valley fill. The Turtle Creek soil, which is named for an exposure in a bluff along Turtle Creek in the SW¼ SW¼ sec. 27, T. 20 N., R. 15 W., Valley County, Nebraska, was developed on this alluvium during the climatic optimum. Late Recent includes deposits younger than the climatic optimum. Unconformably overlying the Turtle Creek soil and at the top of the high silt terrace is silty alluvium on which is developed the Arcadia soil. The Arcadia soil is named for an exposure in a deep gully southwest of the middle Loup River in the NE¼ NW¼ sec. 34, T. 17 N., R. 16 W., Valley County, Nebraska. Recent fill 1, a silty sandy alluvium in a terrace about 15 feet below the high silt terrace, also has the Arcadia soil developed on it. Recent fill 2, a humic sandy alluvium, lies in a terrace 11 feet below the surface of Recent fill 1. No soil is developed on Recent fill 2.

Terraces

Only three terraces occur along the streams in the Valley County area. The oldest terrace (high silt terrace) is more than 30 feet above stream level. It contains deposits ranging in age from Kansan to Recent. The next-younger terrace (terrace surface 1) is about 22 feet above stream level. It contains only Recent fill 1. The youngest terrace (terrace surface 2) is about 11 feet above stream level. It contains only Recent fill 2.

The oldest terrace is 1–3 miles wide and extends along most large streams in the area. The two younger terraces are small fill terraces in arroyos cut in the high silt terrace or in the loess.

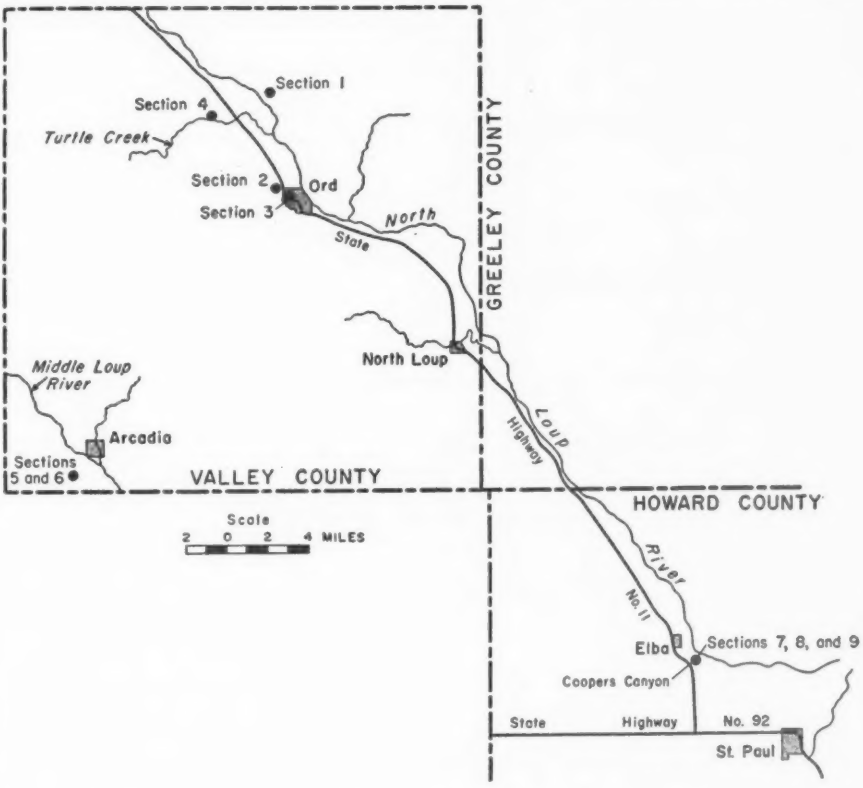

FIGURE 2.—LOCATION OF MEASURED SECTIONS IN TERRACES OF WISCONSIN AND RECENT AGES

DESCRIPTIONS OF SPECIFIC SECTIONS

The sequence of alluviation was examined at 15 localities, and six localities including nine sections were selected for discussion (Fig. 2). These nine are shown in diagrammatic form on Figures 3 and 4, and a detailed measurement of section 7 is included.

SECTION 1: Section 1 (Fig. 3) was measured in a roadside ditch on the northeast side of the North Loup River, in the NE¼ NW¼ sec. 29, T. 20 N., R. 14 W., near the break in slope between the loess-covered uplands and the terrace. Tazewell(?), Cary(?), and Mankato(?) to early Recent (undifferentiated), and late Recent alluvium are exposed, and the soil sequence encompasses all soils from the Brady(?) soil to the Arcadia soil except the Elba soil. Though the deposits are thin and the soils are nearly superposed, good profiles are developed and lime is abundant in the Brady(?) and Ord soils.

SECTION 2: Section 2 (Fig. 3) was measured along the Ord-Sargent road in the SW¼ SE¼ SE¼ sec. 17, T. 19 N., R. 14 W. A terrace remnant is cut by the road and because the Man-

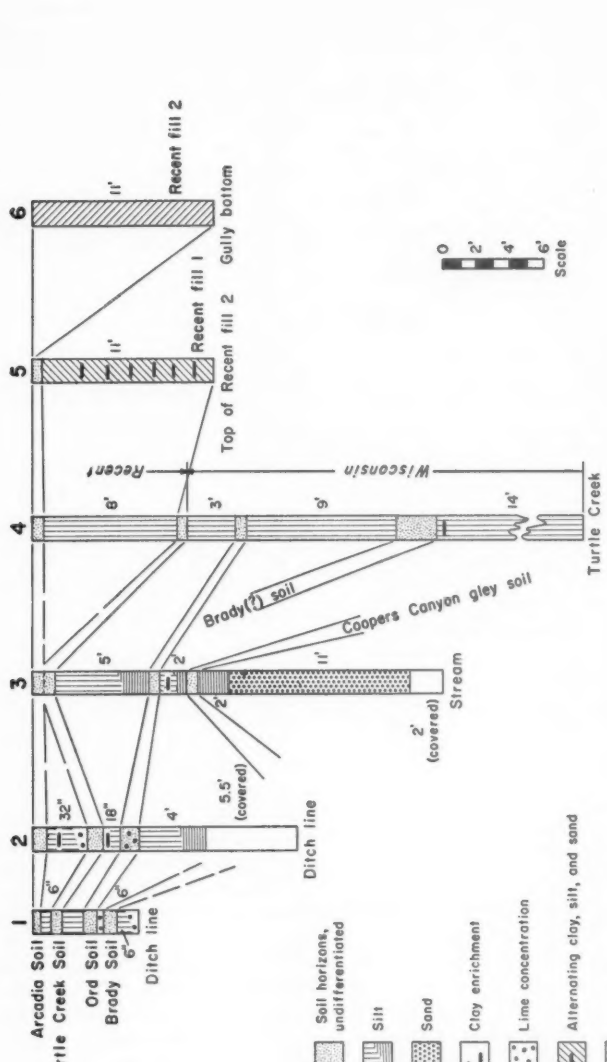

FIGURE 3.—CORRELATION OF WISCONSIN SOILS AND RECENT FILLS
Along the North Loup River and its tributaries, Valley County area, Nebraska.

kato(?) and early Recent (undifferentiated) and late Recent alluvial deposits are thin the Ord, Turtle Creek, and Arcadia soils are exposed in near superposition. All three soils are in the

Physa anatina Lea (mostly young)
P. elliptica Lea
Gastrocopta armifera (Say) (broken)
G. tappaniana (C. B. Adams)
Vertigo ovata Say

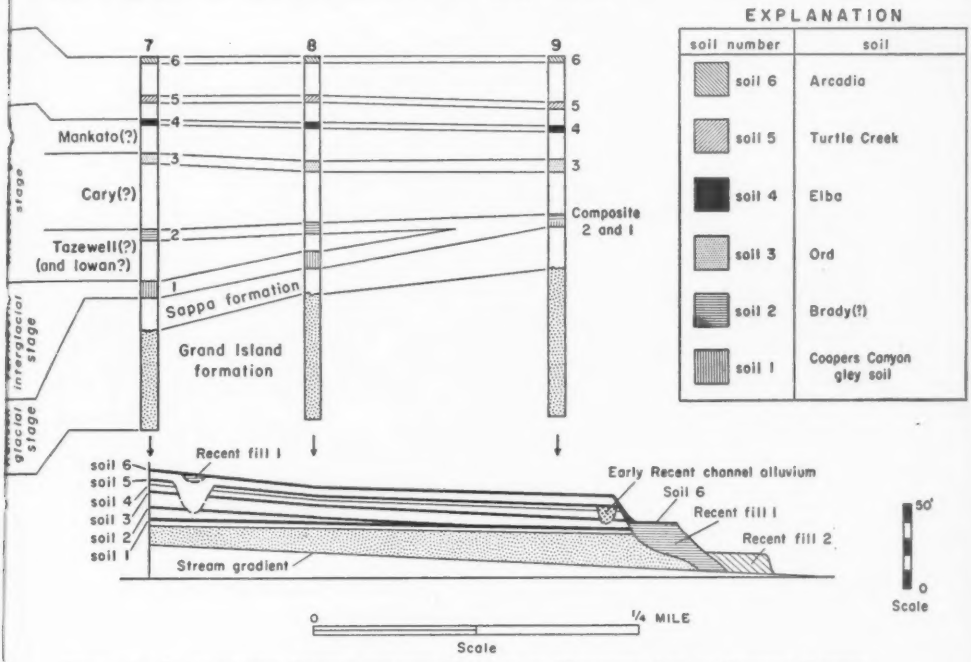

FIGURE 4.—DIAGRAMMATIC SECTION THROUGH TERRACE OF WISCONSIN AGE AT COOPERS CANYON

upper 10.5 feet of a 16-foot exposure. The Cary(?) silt below the Ord soils is slightly stratified in the lower part and very fossiliferous. The 31 species listed were collected by the authors and identified by Joseph P. E. Morrison. The molluscs indicate a floodplain environment. None of them is restricted to a particular substage of the Wisconsin stage.

Pisidium compressum Prime
P. neglectum Sterki (?) (young)
P. rotundatum Prime (?)
Valvata sp. (too young)
V. tricarinata (Say)
Stagnicola caperata (Say)
S. palustris elodes (Say)
Fossaria parva (Lea)
F. modicella (Say) (?)
Gyraulus deflectus (Say) (?) (young)
G. circumstriatus (Tryon)
G. parvus (Say)
Helisoma anceps (Menke)
Promenetus exacuous (Say) (young)
P. umbilicatellus (Cockerell) (??) (young)
V. modesta (Say)
V. milium (Gould)
Vallonia cyclophorella Ancey
V. perspectiva Sterki
Helicodiscus parallelus (Say)
Succinea avara Say (mostly young)
S. sp. (too young)
Discus cronkhitei (Newcomb)
Zonitoides arboreus (Say)
Punctum minutissimum (Lea)
Sphaerium simile (Say) (fragment)

SECTION 3: Section 3 (Fig. 3) was measured in a nearly vertical embankment on the west side of the golf course at Ord (SW¼ NW¼ sec. 21, T. 19 N., R. 14 W.). Three soils are exposed; the lower soil is correlated with the Coopers Canyon gley soil. The following fossils from the Coopers Canyon gley soil at this locality were identified by Dwight W. Taylor:

Pisidium compressum Prime
P. casertanum (Poli)
Valvata tricarinata (Say)

Carychium exiguum (Say)
Lymnaea caperata Say
L. parva Lea
Gyraulus parvus (Say)
G. circumstriatus (Tryon)
Armiger crista (Linné)

the Grand Island formation A stratified compact dark-gray silt underlying the sand is considered to be a lens in the Grand Island formation. The Brady(?) soil is absent. The soil above

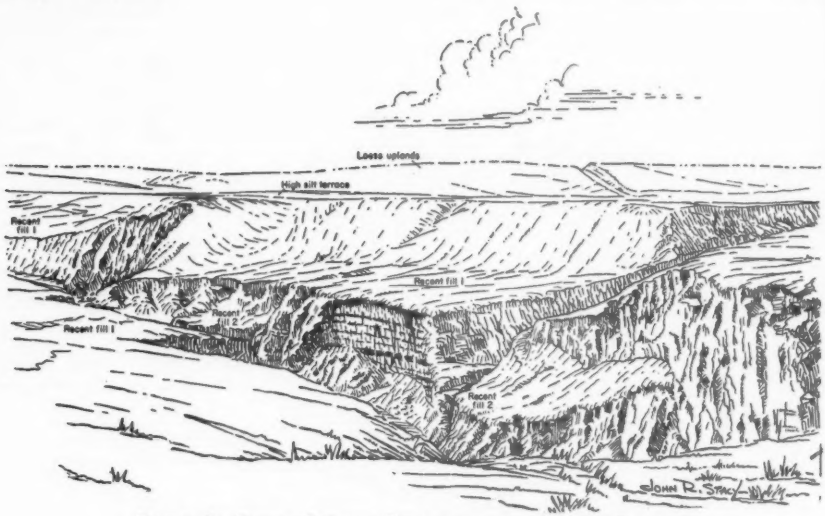

FIGURE 5.—SKETCH SHOWING RELATIONSHIP OF TERRACE SURFACES

Helisoma sp.
Promenetus pearlettei (Leonard)
Physa skinneri Taylor
Aplexa hypnorum (Linné)
Ferrissia sp.
Gastrocopta tappaniana (C. B. Adams)
G. contracta (Say)
Vertigo tridentata Wolf
V. ovata Say
V. milium (Gould)
Vallonia gracilicosta Reinhardt
V. cyclophorella Sterki
Succinea sp.
Discus cronkhitei (Newcomb)
Punctum minutissimum (Lea)
Euconulus fulvus (Müller)
Retinella electrina (Gould)
Strobilops sparsicostata (Baker)
Hawaiia minuscula (Binney)
Helicodiscus parallelus (Say)

Promenetus pearlettei (Leonard) is considered diagnostic of late Kansan and Yarmouth deposits according to Dwight W. Taylor (Personal communication) and Frye and Leonard (1952, p. 153).

The silt below the Coopers Canyon gley soil lies unconformably upon a white medium-grained sand, which contains lenses of coarser sand and gravel. The sand was correlated as the Coopers Canyon gley soil is the Ord soil (type locality). The Ord soil has an A horizon 6–10 inches thick and a B horizon 6 inches thick and overlies 2 feet of slightly calcareous silt. The youngest soil in this exposure has been correlated as a composite of the Turtle Creek and Arcadia soils because of its thickness and because of the intensity of clay development in its B horizon. The silt below this composite soil is an excellent example of the alluvial equivalent of the upland loess. The lower part of the silt is stratified and the upper part is massive.

SECTION 4: Section 4 (Fig. 3) was measured in a bluff along Turtle Creek (SW¼ SW¼ sec. 27, T. 20 N., R. 15 W.), one of the major tributaries of the North Loup River in Valley County. Terraces along this tributary, and along most others in the area, display the same sequence of alluviation and soil formation as do the terraces of the Middle Loup and North Loup rivers. The Turtle Creek soil is named from this exposure. Its A horizon is 6–8 inches thick and its B horizon is moderately clay-

enriched. The silts are massive with columnar joints.

SECTIONS 5 AND 6: Sections 5 and 6 were measured southwest of the Middle Loup River in the NE¼ NW¼ sec. 34, T. 17 N., R. 16 W., in a deep gully in which are exposed two late Recent fills, the older of which is designated Recent fill 1, and the younger Recent fill 2. The two fills are persistent, are easily recognized by their terrace surfaces, and may be differentiated by their height above stream level (Fig. 5).

Recent fill 1 fills a channel in the older deposits of the high silt terrace at the head of the gully, and is composed of tan clayey silt layers 16–18 inches thick separated by humic layers 2–4 inches thick. The Arcadia soil is formed on Recent fill 1 and is named from this exposure. Its A horizon is 4 inches thick and contains some lime, and its B horizon is 2 inches thick and only weakly clay-enriched.

Recent fill 2 is composed of layers of tan silt and gray fine sand, and contains less clay than Recent fill 1. Commonly, limonite stains the joint planes of this fill. The terrace level of Recent fill 2 is 11 feet lower than that of Recent fill 1, and the alluvium generally is exposed in vertical cut banks in the gully.

SECTIONS 7, 8, AND 9: Sections 7, 8, and 9 were measured at widely spaced intervals from west to east along Coopers Canyon, which is cut into the terrace of the North Loup River near Elba in Howard County, in sec. 14, T. 15 N., R. 11 W. In Coopers Canyon, terrace deposits are exposed almost continuously for more than half a mile in a direction transverse to the river. The Grand Island formation exposed at the base of the terrace is overlain by the Sappa formation (Biemesderfer, in preparation). A greater thickness of the Grand Island formation is exposed near section 9 at the mouth of the canyon than upstream at section 7. A well-developed gley soil, called the Coopers Canyon gley soil, with a humic zone 24 inches thick is formed on the Sappa formation.

The Tazewell(?) alluvial silt between the Coopers Canyon gley soil and Brady(?) soil thins rapidly toward the center of the North Loup River valley where the soils are superposed upon one another and form a single thick strongly developed composite soil (sec. 9 and Fig. 4). The Cary(?) alluvium between the Brady(?) and Ord soils also decreases in thickness toward the center of the valley where the Brady(?) and Ord soils probably are also superposed. The Mankato(?) silty alluvium overlies

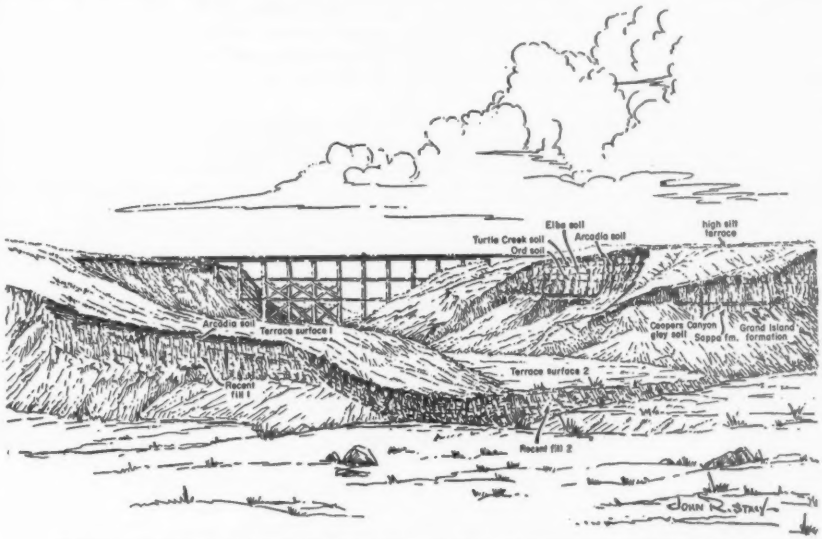

FIGURE 6.—SKETCH OF TERRACE DEPOSITS AND SOIL HORIZONS AT COOPERS CANYON

TABLE 2.—SECTION 7

Fifty yards east of bridge on State Highway 11, in Coopers Canyon, sec. 14, T. 15 N., R. 11 W., Howard County, Nebraska.

Soil name	Soil horizon	Description*	Thickness (feet)	Depth (feet)
		Vegetation: short grass.		
Arcadia	A	Dark-gray-brown (10 YR 4/2) humic, silty clay loam. *Structure*: fine weak platy; *consistence*: sticky, slightly plastic, loose; *reaction*: slightly acid. Diffuse boundary†	2.2	0–2.2
	B	Gray-brown (2.5 Y 5/2) silty clay loam. *Structure*: fine weak prismatic; *consistence*: sticky, plastic, soft; *reaction*: strongly acid; Gradual boundary	1.0	2.2–3.2
	C_{ca}**	Light-gray (2.5 Y 7/2) silt with veinlets of $CaCO_3$. *Structure*: medium weak columnar; *consistence*: slightly sticky, plastic, soft; *reaction*: strongly alkaline. Clear boundary	2.2	3.2–5.4
Turtle Creek	A	Light-gray (10 YR 7/2) humic silt loam. *Structure*: fine weak granular; *consistence*: sticky, plastic, slightly hard; *reaction*: medium alkaline, secondary line. Shell fragments. Gradual boundary	2.0	5.4–7.4
	B	Light-gray (10 YR 7/2) silt. *Structure*: fine weak columnar; *consistence*: slightly sticky, plastic, slightly hard; *reaction*: medium alkaline, secondary lime. Weak to strong stratification in channels. Gradual to abrupt boundary. Topography of boundary†† smooth to broken	1.0	7.4–8.4
Elba	A	Pale-brown (10 YR 6/3) humic silt loam. *Structure*: fine weak granular; *consistence*: sticky, plastic, slightly hard; *reaction*: medium alkaline, secondary lime. Gradual boundary	1.5	8.4–9.9
	B	Light-gray (10 YR 7/2) silt. *Structure*: medium strong columnar; *consistence*: slightly sticky, plastic, hard; *reaction*: medium alkaline, secondary lime. Slightly fossiliferous. Gradual boundary	1.0	9.9–10.9
Ord	A	Light brownish-gray (10 YR 6/2) humic silt loam with stringy lime veinlets. *Structure*: medium moderate prismatic; *consistence*: slightly sticky, plastic, very hard; *reaction*: medium alkaline, secondary lime. Diffuse boundary	1.8	10.9–12.7
	B	Light-gray (10 YR 7/2) silt. *Structure*: columnar; *consistence*: sticky, plastic, soft; *reaction*: strongly alkaline, secondary lime. Diffuse boundary	2.0	12.7–14.7

TABLE 2.—SECTION 7—*Cont.*

Soil name	Soil horizon	Description*	Thickness (feet)	Depth (feet)
	C_{ca}	Light-gray (10 YR 7/2) silt. *Structure*: columnar; *consistence*: slightly sticky, plastic, soft; *reaction*: strongly alkaline. Diffuse boundary...	3.0	14.7–17.7
	C_{ca}	Light-gray (10 YR 7/2) silt with thin-walled lime concretions (one-eighth inch in diameter). *Structure*: massive; *consistence*: slightly sticky, plastic, soft; *reaction*: strongly alkaline. Clear boundary.....	2.0	17.7–19.7
Brady(?)	A	Grayish-brown (10 YR 5/2) humic silty loam with platy nodular $CaCO_3$ concretions. *Structure*: fine weak columnar; *consistence*: slightly sticky, plastic, soft; *reaction*: strongly alkaline, secondary lime. Slightly fossiliferous. Clear boundary.....	1.3	19.7–21.0
	B	Light-gray (2.5 Y 7/2) silt. *Structure*: massive; *consistence*: slightly sticky, plastic, soft; *reaction*: slightly alkaline, secondary lime. Diffuse boundary.....	.7	21.0–21.7
	C_{ca}	Light-gray (2.5 Y 7/2) silt with nodular lime concretions and small channels of coarse sand composed of one-eighth inch diameter limonite pipettes. *Structure*: massive; *consistence*: slightly sticky, slightly plastic, soft; *reaction*: slightly alkaline. Diffuse boundary between alternating layers of silt and humic silt within layer. Slightly fossiliferous. Diffuse boundary...	1.0	21.7–22.7
	C_{ca}	Light-brownish-gray (10 YR 6/2) silt. *Structure*: coarse strong columnar; *consistence*: slightly sticky, slightly plastic, slightly hard; *reaction*: medium alkaline. Gradual boundary..	1.3	22.7–24.0
Sappa formation: Coopers Canyon gley soil	G	Gray (10 YR 5/1) humic silt loam. *Structure*: coarse moderate columnar; *consistence*: slightly sticky, slightly plastic, slightly hard; *cementation*: weakly cemented; *reaction*: medium alkaline, secondary lime. Slightly fossiliferous. Diffuse boundary.....	1.2	24.0–25.2
	G	Gray (10 YR 6/1) humic silt loam. *Structure*: fine moderate granular; *consistence*: slightly sticky, slightly plastic, slightly hard; *cementation*: strongly cemented; *reaction*: strongly alkaline, secondary lime as nodular lime concretions one-fifth inch maximum diameter. Fossils extremely abundant. Clear boundary..	.8	25.2–26.0
	C	Pale-yellow (5 Y 8/3) stratified silt with reddish yellow (7.4 YR 6/8) limonite stain. *Structure*: coarse strong columnar; *consistence*: slightly sticky, slightly plastic, hard; *cementation*: weakly cemented; *reaction*: slightly alkaline, secondary lime. Slightly fossiliferous. Clear boundary.....	1.2	26.0–27.2

TABLE 2.—SECTION 7—*Cont.*

Soil name	Soil horizon	Description*	Thickness (feet)	Depth (feet)
Grand Island formation		Covered. White (7.5 YR 9/0) slightly cross-bedded medium-grained quartz sand is exposed in places.....	12.0	27.2–39.2
		Bottom of gully		

* Color names and designations refer to the rock color chart of the National Research Council.

† Distinctness of boundaries: abrupt, less than 1 inch; clear, 1 inch to 2½ inches; gradual, 2½ inches to 5 inches; diffuse, plus 5 inches.

** The subscript ca refers to a layer of accumulated calcium carbonate.

†† Topography of boundary: smooth, nearly a plane; wavy, contains pockets wider than their depth; irregular, contains pockets deeper than their width; broken, parts of the horizon are unconnected with other parts: (U. S. Dept. of Agriculture, 1951, p. 187).

the Ord soil, and on the alluvium is developed the Elba soil which is named for this exposure. The Elba soil has an A horizon 1.5 feet thick (sec. 7). Early Recent silty alluvium overlies the Elba soil. A small channel was cut and filled prior to the development of the Turtle Creek soil which is developed on the Early Recent alluvium and the channel fill. The youngest conformable deposit in section 7 is late Recent silty alluvium. On this alluvium the Arcadia soil is developed as the uppermost soil on the high silt terrace (Fig. 6) and on Recent fill 1. A small channel cut near the valley wall down through the Brady(?) soil is filled with this late Recent alluvium. Recent fill 1, a silty sandy alluvium, is deposited in Coopers Canyon which is cut through the high silt terrace. Recent fill 2, a humic sandy alluvium, is deposited in an arroyo cut through Recent fill 1.

Section 7 (Table 2) is typical of all three sections in Coopers Canyon.

Molluscs collected from arbitrarily chosen zones of the Coopers Canyon gley soil (*see* Fig. 7) were identified by Dwight W. Taylor. The fossils were collected from section 9 along Coopers Canyon in sec. 14, T. 15 N., R. 11 W., southeast of Elba, Nebraska.

Molluscs obtained from an unsorted sample of the Coopers Canyon gley soil were identified by the authors. Joseph P. E. Morrison checked the authors' identifications and made ecological interpretations. The following fauna was collected from the lower part of the Coopers Canyon gley soil in section 9 (Fig. 4) along Coopers Canyon southeast of Elba, Nebraska (sec. 14, T. 15 N., R. 11 W.).

Pisidium lilljeborgi Clessin
P. rotundatum Prime (?)
P. compressum Prime
P. neglectum Sterki
Valvata tricarinata (Say)
V. sincera (Say)
Amnicola limosa porata (Say)
A. walkeri Pilsbry
Carychium exiguum (Say)
Lymnaea stagnalis jugularis Say
Stagnicola caperata (Say)
S. palustris elodes (Say)
Fossaria modicella (Say)
F. exigua (Lea)?
F. parva (Lea)
Gyraulus similaris F. C. Baker
G. circumstriatus (Tryon)
G. deflectus (Say)
G. deflectus (Say) (young)
Armiger crista (Linnaeus)
Helisoma anceps (Menke)
H. trivolvis (Say)
Promenetus exacuous (Say)
Physa eliptica Lea (young)
P. anatina Lea
P. halei Lea
Aplexa hypnorum (Linnaeus)
Ferrissia parallela (Haldeman)
Gastrocopta tappaniana (C. B. Adams)
G. contracta (Say)
Vertigo milium (Gould)
V. morsei Sterki
V. ovata Say
V. elatior Sterki
Vallonia cyclophorella Ancey
Succinea avara gelida F. C. Baker
S. grosvenori Lea

	1	2	3	4	
Pisidium casertanum (Poli)		X	X		
P. compressum Prime			X		
P. obtusale Pfeiffer	X	X	X		
Valvata tricarinata (Say)			X		**
Amnicola limosa (Say)			X		
A. walkeri Pilsbry			X		
Carychium exiguum (Say)	X	X	X		
Lymnaea caperata Say	X	X	X		
L. obrussa Say			X		
L. palustris (Muller)			X		
L. parva Lea	X		X	sp.	
Gyraulus circumstriatus (Tryon)	X	X	X		
G. hirsutus (Gould)			X	X	**
G. parvus (Say)			X	X	
Armiger crista (Linne)			X		*
Helisoma trivolvis (Say)			X		
Promenetus pearlettei (Leonard)			X	X	
P. umbilicatellus (Cockerell)	X	X	X		
Physa anatina Lea			X	X	
P. elliptica Lea			X		
P. skinneri Taylor			X		
Aplexa hypnorum (Linne)		X	X		*
Ferrissia sp.			X		
Gastrocopta contracta (Say)	X		X	X	
G. tappaniana (C. B. Adams)	X	X	X	X	
G. armifera (Say)		?			
Vertigo gouldi basidena Pilsbry and Vanatta			X		
V. milium (Gould)		X	X		
V. ovata Say	X	X	X	X	
Vallonia perspectiva Sterki				X	
V. sp.			X		
Succinea cf. ovalis Say			X		
S. sp.	X		X	X	
S. cf. avara Say			X		
Oxyloma haydeni (Binney)			X		
Discus cronkhitei (Newcomb)			X	X	
Punctum minutissimum (Lea)			X		
Euconulus sp.		X	X		
Retinella electrina (Gould)			X	X	
R. indentata (Say)			X		
Pupoides albilabris (C. B. Adams)	X				
Strobilops sp.	X				
Deroceras aenigma Leonard	X	sp.			
Hawaiia minuscula (Binney)	X	X		X	
Sphaerium sp.		X			
Planorbula armigera (Say)		X			

** markedly northern habitat

* northern habitat

FIGURE 7.—DISTRIBUTION OF MOLLUSC FROM ZONES IN COOPERS CANYON GLEY SOIL
Each zone represents one-fourth the thickness of the gley soil, Coopers Canyon, Howard County, Nebraska; zone 1 is at the top

S. ovalis Say
Oxyloma haydeni (W. G. Binney)
O. decampi peoriensis (Walker) (?)
Discus cronkhitei (Newcomb)
Euconulus fulvus (Müller)
Retinella electrina (Gould)
Strobilops labyrinthica (Say)
S. aenea mexicana Pilsbry and Ferriss
Zonitoides arboreus (Say)
Deroceras cf. *laeve* (Müller)
Hawaiia minuscula (Binney)
Sphaerium simile (Say)
Planorbula armigera palustris F. C. Baker

The Coopers Canyon fauna required quiet water—lakes, ponds, swamps, slow-moving clear streams, wet grassy margins of streams, or moist woodlands.

Promenetus pearlettei (Leonard) is considered diagnostic of late Kansan to Yarmouth deposits (D. W. Taylor, personal communication; Frye and Leonard, 1952, p. 153).

Fossils collected from this locality by G. K. Biemesderfer were identified by A. Byron Leonard who stated that the assemblage is a characteristic Sappa fauna (Biemesderfer, in preparation).

Basis for Assigning Ages

On the basis of lithological characteristics, stratigraphy, and fossils, G. K. Biemesderfer has established the identity of the Sappa formation in Howard County and Miller the identity of the Sappa formation in Valley County.

The Brady(?) soil of the valleys is correlated with the Brady soil of the uplands on the basis of its development. The A horizon of each is slightly more than a foot thick, is a grayish-brown, humic silty loam with fine weak columnar to coarse weak columnar structure, a slightly sticky, plastic, soft consistence, and pH strongly alkaline from secondary lime. The B horizon of each is about 8 inches thick, is light gray and silty, with massive to coarse strong columnar structure, slightly sticky, plastic, soft consistence, and pH slightly alkaline from secondary lime.

Thorp *et al.* (1951, p. 14) describe the Brady soil 1½ miles south of Bignell, Lincoln County, Nebraska as follows:

"11 ft. 6 in.–12 ft. 8 in. (beneath the surface) A horizon of Brady soil. Dark greyish-brown (dry), to very dark greyish-brown (moist) weak-prismatic, friable silt loam that breaks easily to weak subangular blocks. Contains calcium carbonate in threads and films, leached down from material above.

"12 ft. 8 in.–15 ft. B horizon of Brady soil. Pale brown (dry), to dark greyish-brown (moist), weak prismatic silt loam that breaks easily to weak subangular blocks. Contains numerous threads and films of calcium carbonate leached down from loess above."

The very weak Arcadia soil is the youngest soil in the area and is late Recent in age.

The moderately strong Turtle Creek soil (formed during the climatic optimum) is only slightly weaker than the soil developed on the Bignell loess. The soil on the Bignell loess is stronger because it is the result of the soil-forming processes of both the Turtle Creek and the Arcadia times, and locally may also be the result of the soil-forming process of Elba time.

The other Wisconsin and Recent deposits and soils in the valleys lie between the strong Brady(?) soil and the very weak Arcadia soil. Three beds of alluvium and two soils lie above the Brady(?) soil and below the Arcadia soil.

The very weak Elba soil was observed only in Coopers Canyon. Because of its limited areal extent its correlation with a soil-forming time at the end of the Mankato substage is even more tenuous than the other correlations. The Elba soil is potentially important to the interpretation of the history of the Mankato substage, and suggests a definite break between the Pleistocene and Recent deposits in the Great Plains.

The weak Ord soil lies between the Brady(?) soil and the Elba soil and probably formed in the interglacial between the Cary and Mankato substages. Its great thickness compared to its weakness suggests that the climate of this soil-forming time was not as stable as the climate of the soil-forming times of the Brady(?) soil and the Turtle Creek soil. In other words loess and alluvium were being deposited as the soil was forming.

Geologic History

Only events of Kansan and later ages will be discussed under geologic history.

The sand of the Grand Island formation probably was deposited on wide sandy flood plains. As the available water diminished in late Kansan time, the valleys were only inter-

mittently flooded. Many ponds and swamps lay on the surface of the flood plain and the floods laid down the silt of the Sappa formation. Further proof of a flood plain environment is provided by the molluscs which require slow-moving clear streams, lakes, or ponds. A similar environment probably attended the development of the Coopers Canyon gley soil on the Sappa silt during the Yarmouth interglacial stage. The absence of the soil and part of the Sappa formation at all but three localities indicates post-Yarmouth erosion. The Crete formation of Illinoian age probably is a deposit of this erosional interval. The Loveland loess accumulated in the dissected uplands later in Illinoian time.

The reddish-brown Sangamon soil developed on the Loveland loess in Sangamon time.

In the valleys no identifiable Loveland loess or Sangamon soil was observed. One of three hypotheses may explain the absence of the Loveland loess and the Sangamon soil in the valleys:

(1) Deposition of Loveland loess, but locally in such small amounts that it is part of the Coopers Canyon gley soil.

(2) Deposition and subsequent removal of the loess.

(3) No deposition in the valleys during Illinoian and Sangamon times.

For the following reasons the first explanation seems most logical, but locally either of the other explanations may be possible.

(1) The Sangamon soil is developed on pre-Illinoian deposits at localities observed by the authors in eastern Nebraska, eastern Kansas, and western Iowa. Either the Loveland loess was not deposited at these localities or it was deposited in such small amounts that it appears to be part of the pre-existing deposit. Similarly, in Valley County, although Peorian loess ranges in thickness from 15 to 25 feet in the upland, the alluvial silt equivalent in the valleys is everywhere less than 10 feet in thickness.

(2) Erosional features—such as irregularities on the surface of the soil, removal of parts of the soil profile, or channels filled with younger alluvium in either the Coopers Canyon gley soil or the Sappa formation—that would indicate removal of the Loveland loess are lacking.

Lack of these erosional features would probably eliminate the possibility of any of the following sequences of events:

(1) deposition of Sappa formation
erosion of part of the Sappa formation
development of Yarmouth soil on thinned Sappa
development of Sangamon soil

(2) deposition of Sappa formation
erosion of part of Sappa
development of Yarmouth soil
deposition of Loveland loess
erosion of Loveland loess and part of Yarmouth soil or only Loveland loess
development of Sangamon soil

(3) deposition of Sappa formation
development of Yarmouth soil
erosion of part of Yarmouth soil
development of Sangamon soil

(4) deposition of Sappa formation
development of Yarmouth soil
deposition of Loveland loess
erosion of Loveland loess with or without part of Yarmouth soil
development of Sangamon soil.

Any one of these sequences that involve erosion of the Sappa formation would seem improbable because it would require a thicker deposit of Sappa formation than has been observed by the writers in this part of Nebraska.

A detailed study of the Sappa formation and Coopers Canyon gley soil at section 7 in Coopers Canyon reveals a history of apparently continuous deposition. Molluscs first appear only slightly above the base of the Sappa. Because the lithological characteristics of the lower 0.8 foot of the Coopers Canyon gley soil are the same as the parent Sappa formation the soil probably was developed on Sappa. *Promenetus pearlettei* (Leonard), which is diagnostic of late Kansan and Yarmouth deposits, is found in the lower 0.8 foot of the Coopers Canyon gley soil. The upper 1.2 foot of the Coopers Canyon gley soil is slightly more silty than the lower part and does not contain *Promenetus pearlettei* (Leonard). The absence of this diagnostic form from the upper part of the Coopers Canyon gley soil may indicate that the upper part is of later age, probably Sangamon. That part of the soil formed during Yarmouth time contains molluscs that require fresh, quiet water; that part formed in post-Yarmouth (probably Sangamon) time also contains molluscs that require fresh, quiet water. The floodplain probably was almost continuously saturated and locally covered by ponds, swamps, and abandoned channels.

According to Joffe (1949, p. 419)

"Glei formation is typical for marshes and conditions of a high water table. The waterlogged state of the glei horizon (G) brings about anaerobic conditions which favor reducing reactions and minimizes leaching effects. Besides, the ground waters rise by capillary forces and enrich the glei horizon with bases, if such are in solution, and impart to it an alkaline reaction. In the reduction process, the ferric iron gives rise to the ferrous iron, and in the case of the hydroxide, the weak base changes into the stronger ferrous iron base. One should, therefore, expect a higher pH in the glei horizon, party (sic.) because of this."

This reducing environment resulted in the lack of a reddish-brown coloring, common to most Sangamon soils in Nebraska, in the upper part of the Coopers Canyon gley soil.

After development of the Sangamon soil in the uplands and the Sangamon gley component of the Coopers Canyon gley soil in the valleys, loess blanketed the uplands. The earliest Wisconsin loess in the uplands is the Peorian loess of Iowan and Tazewell age. The Brady soil is developed on the Peorian loess, and a soil correlated with the Brady soil is developed on the lowest bed of Wisconsin—Tazewell(?)—alluvial silt in the valleys. At section 7 in Coopers Canyon 4.3 feet of Tazewell(?) alluvium with the Brady(?) soil developed on it overlies the Coopers Canyon gley soil. At section 9 the Brady(?) soil is developed on a slightly thickened Coopers Canyon gley soil. The slight increase in thickness of the Coopers Canyon gley soil at section 9 is attributed to the addition of a small amount of Tazewell(?) alluvium. The superposition of the Brady(?) soil on the Coopers Canyon gley soil at section 9 poses the problem of accounting for the absence of about 4 feet of Tazewell(?) alluvium.

In Figure 4 the top of the Coopers Canyon gley soil is shown as a line sloping gently eastward, whereas the younger soils are shown as lines that dip more steeply eastward toward the river. The wedgeout of Tazewell(?) alluvium can be traced along Coopers Canyon. The Tazewell(?) alluvium may have been eroded to produce the wedgeout shown in Figure 4 but it seems doubtful because erosion probably would have removed part of the Coopers Canyon gley soil beyond where the Tazewell(?) alluvium wedges out. One of the following possibilities may explain the thinning of Tazewell(?) alluvium and consequent convergence of the Brady(?) soil and the Coopers Canyon gley soil.

(1) Slopewash accumulated along the sides of the valley, but there was very little or no deposition in the center of the valley.

(2) Larger streams removed loess as fast as it was deposited in the valleys. Smaller streams in the first stage removed the loess as it was deposited; in the second stage only reworked the loess into alluvial silt as it clogged the valleys; and in the final stage were completely choked and the loess blanketed the alluvial silt as shown in sections 3 and 4, Figure 3. As a result, silt accumulated along the valley sides of the larger streams owing to the incompetency of the tributaries to remove the silt.

The absence of loess in the larger valleys or thick deposits of alluvial silt derived from loess may be further explained by the likelihood that the large valleys were the local sources of loess. Turbulence over these larger valleys probably prevented deposition of wind-borne silt. In any event, although Peorian loess ranges in thickness from 15 to 25 feet in the upland, the alluvial-silt equivalent in the valleys is everywhere less than 10 feet thick.

Bignell loess of Cary and Mankato ages was deposited on the Peorian loess. Locally the Brady soil is preserved and the Bignell loess overlies it.

Cary(?) alluvium is deposited on the Brady(?) soil in the valleys. The Ord soil is developed on the Cary(?) alluvium. The Mankato(?) alluvium is deposited on the Ord soil. The Elba soil developed on Mankato(?) alluvium at the end of Mankato(?) time. Early Recent silty alluvium was deposited on the Elba soil locally, but generally is deposited directly on Mankato(?) alluvium and is indistinguishable from it. Channels were eroded into underlying deposits at one place through the Ord soil. The Turtle Creek soil is developed on the early Recent alluvium.

Eolian sand and loess were deposited on the uplands in early Recent time.

In the valleys late Recent alluvium was deposited on the Turtle Creek soil and is the upper alluvial bed of the high silt terrace. During late

Recent time some channels were cut through the Coopers Canyon gley soil. Recent fill 1 was deposited in these channels, and the Arcadia soil developed on both the Recent fill 1 and on the late Recent upper alluvium of the high silt terrace. Recent fill 1 was removed in places, and Recent fill 2 was deposited in these later channels. Later erosion re-excavated Recent fill 2 to the level of the modern gully floor.

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Short Notes

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ABUNDANCE OF Sr^{87} DURING GEOLOGIC TIME

By Paul W. Gast

Variations in isotope abundances due to the addition of radiogenic isotopes during geologic time may yield important information about the earth's crust. This has been demonstrated in the case of radiogenic lead derived from the decay of uranium and thorium. Wickman (1948) suggested that the radioactivity of Rb^{87} might be a cause of measurable changes in the isotope abundance of Sr^{87}, in rocks and minerals, during geologic time. In particular he proposed that the increase in the Sr^{87} abundance be used to determine the absolute age of marine carbonates and shells. Such materials are particularly suitable for the study of this variation because there is a good separation of Rb and Sr in calcium carbonate minerals. Provided the strontium in the oceans represents an average isotope abundance for Sr in surface rocks, the Sr^{87} isotope abundance in these materials should decrease almost linearly with their age.

Early work on this problem at the Lamont Observatory (J. L. Kulp, personal communication) suggested that the variation was smaller than that expected by Wickman. The published work of Herzog *et al.* (1953) also suggested that the change in abundance was much less than that predicted, but the uncertainty in the age of the samples and in the isotopic analyses due to isotope fractionation during measurement (Herzog, 1954) make it difficult to evaluate accurately how much the Sr^{87} abundance has changed with time. Thus this further study was initiated with two objectives in mind: (1) to set limits on the possible application of the proposed strontium isotope method for measuring age of sedimentary rocks; (2) to determine if other types of geologic information can be derived from study of Sr isotopes.

Experimental Problem

One half to one gram of limestone was dissolved in concentrated HCl; the insoluble residue was filtered off. All samples described here were more than 90% soluble in HCl. Separation of Sr from Ca was effected by an ion-exchange process. The chloride solution was absorbed on a 2.2 by 20 cm bed of Dowex 50–8%, 200–400 mesh and eluted with 3N HCl. It is estimated that 400–800 micrograms of Sr was separated from a gram sample.

The mass spectrometer used to determine isotopic abundances has been described by Aldrich *et al.* (1953). However, in the work reported here the isotope abundances were determined using a single collector and magnetic sweep scanning instead of the double-collector technique previously used. Such scans were made in sets of 10 to 20; the atom ratios Sr^{86}/Sr^{88}, Sr^{87}/Sr^{88}, and Sr^{87}/Sr^{86} were determined for each scan. The mean deviation on a set of scans was usually less than 0.5 per cent. Both plain and nickel-plated tungsten ribbon were used for filaments. Twenty five to 50 micrograms of Sr was used in each run. In most runs data were taken over a period of 8–24 hours, during which time the sample was usually exhausted.

In the course of this work it was observed that during a single run the Sr^{86}/Sr^{88} ratio changed by as much as 4 per cent. Simultaneous changes in the Sr^{87}/Sr^{88} ratio were about half as large, while the changes in the Sr^{84}/Sr^{88} ratio, when determinable, were approximately twice as large as those in the Sr^{86}/Sr^{88} ratio. These variations are attributed to isotope fractionation during evaporation and ionization of Sr from the tungsten ribbon. The direction of the fractionation is such as to de-

plete the sample in the lighter isotopes. Thus, aside from constant systematic discriminations, the highest Sr^{86}/Sr^{88} ratio should be the closest to the true ratio. The highest Sr^{86}/Sr^{88} ratios for the five samples examined in this study are given in Table 1. Each number represents an average of approximately 10 scans with its mean deviation.

TABLE 1.— HIGHEST Sr^{86}/Sr^{88} RATIOS RECORDED

Sample	Highest 86/88 Ratio Recorded
Acantharia shells	.1214 ± .0005
Ordovician limestone	.1210 ± .0005
Belt limestone	.1213 ± .0004
Grenville limestone	.1206 ± .0010
Bulawayan limestone	.1212 ± .0004

Natural variations in the Sr^{86}/Sr^{88} ratio in Sr from celestites, feldspars, and micas have been reported (Aldrich *et al.*, 1953). Later work by Herzog (1954) ascribes at least part of these variations to fractionation during analysis. Data in Table 1 suggest that the Sr^{86}/Sr^{88} ratio is constant within the experimental error for all samples studied here. If it is assumed that the real Sr^{86}/Sr^{88} ratio is constant it can be used to correct the Sr^{87}/Sr^{88} ratio for fractionation which occurs during measurement. If there are real differences in the Sr^{87} content of different samples, these will be seen in the corrected Sr^{87}/Sr^{88} ratio for these samples. Differences in the Sr^{87} content of different samples reported here were found by comparing the corrected Sr^{87}/Sr^{88} and/or Sr^{87}/Sr^{86} ratios. The probable errors given in Table 2 have been estimated from a least-squares analysis of all Sr^{87}/Sr^{88}, Sr^{87}/Sr^{86} and Sr^{86}/Sr^{88} ratios recorded for each sample and the standard. Since absolute values are not known, because of systematic discriminations, data are given in terms of per cent differences from a standard. The standard here is recent celestite ($SrSO_4$) Radiolaria shells (Acantharia) kindly furnished by H. T. Odum, Department of Biology, University of Florida, Gainesville. Isotopic analysis of Sr separated from Mid-Atlantic ocean water shows that the Sr^{87} abundance in the shells and sea water does not differ by more than 0.35 per cent. The differences in the Sr^{87} content found in four different limestones are shown in Table 2. The age of the Ordovician and Beltian samples is inferred from their position in the stratigraphic column. The age given for the Grenville limestone is that suggested by Wilson (1949). The age for the Bulawayan series is a minimum, derived from pegmatites intruding the Bulawayan series (Holmes, 1954).

TABLE 2.—DIFFERENCES IN Sr^{87} ABUNDANCE OBSERVED AND PREDICTED

Sample	Age (M.Y.)	Difference in 87 Abundance (Per cent)	Predicted Variation (Per cent)
Ordovician ls.*	320	+.15 ± 0.40	−1.5
Belt ls.†	650–750	−.35 ± 0.35	−3.2
Grenville ls.	900–1100	−.42 ± 0.40	−5.0
Bulawayan ls.‡	2700	−1.10 ± 0.40	−13.5

* Plattin group, Macy formation, Zell member, Perry County, Texas

† Newland limestone, Belt series, from Walcott collection, Smithsonian Institution

‡ Macgregor (1941)

Calculation of Predicted Variation

The change in Sr^{87} abundance expected from the Rb and Sr content of surface rocks and the Rb^{87} half life is much greater than that observed in these four limestones. The recent measurements of the half life of Rb^{87} vary from values as low as 4×10^{10} years to about 6.3×10^{10} years (Bähnish *et al.*, 1952; Flinta and Eklund, 1954; Bähnish and Huster, 1954). Since even the most recent determinations are in disagreement, an arbitrary value of 5×10^{10} years is used in this paper. This uncertainty in the half life is small compared to other uncertainties involved in this calculation. The abundance of Rb and Sr in surface rocks has been estimated by several investigators. Ahrens *et al.* (1952) have shown that the ratio of K/Rb is constant to within 50 per cent in almost all igneous rocks. Therefore, the abundance of Rb in the exposed land areas can be given in terms of that for potassium, which is

probably accurate to within 25 per cent. Using the K/Rb ratio of 90 given by Ahrens *et al.* (1952) and the estimate of 2.6 per cent as the K content of surface rocks (Rankama and Sahama, 1950, p. 39), the Rb content is found to be 280 g/ton. Goldschmidt *et al.* (1934) made an estimate of 280 g/ton. The abundance of Sr in surface rocks cannot be directly related to a major element; it is therefore difficult to estimate the average Sr content of surface rocks accurately. Goldschmidt (1937) estimates the abundance as 150 g/ton. Noll (1934) in his work on the geochemistry of Sr derives 500 g/ton as the average Sr content of igneous rocks. On the basis of these estimates the arbitrary value of 300 g/ton is used for the present calculation. Thus, the abundance of Rb^{87} is 74 g/ton and of Sr^{87} is 21 g/ton using accepted isotopic abundances. The rate of production of Sr^{87} is 1.0 g/ton per billion years. The change in the abundance of Sr^{87} with time which is derived from these values is 4.8%/b.y. The predicted change in Sr^{87} for the different samples is given in the last column of Table 2. The observed change is less than one tenth of the predicted change.

Discussion

An important conclusion from this data is that the increase of Sr^{87} abundance with time, in calcareous marine sediments and shells, cannot be used as a method of measuring geologic time. Any differences in the Sr^{87} abundance are too small to measure accurately enough with present techniques.

The disagreement between the observed and predicted Sr^{87} abundances is so great that it presents an interesting geochemical problem. Several factors must be considered in any attempt at an explanation.

The first factor is the possibility of extensive ionic exchange of the strontium in the $CaCO_3$ lattice with strontium in circulating ground waters. If carbonate rocks had undergone such exchange to any extent, original variations would be smoothed out. Since such assumed exchange probably would have occurred in recent geologic time, the Sr^{87} abundance would be that of the rock mass adjacent to the limestone.

The rate at which such differences could be destroyed is a function of both the grain size of the $CaCO_3$ containing the Sr and the total amount of Sr which must be exchanged. Thus, the most favorable condition for lack of exchange would be a coarse-grained, extensive $CaCO_3$ deposit. The Grenville marble fulfills these conditions and hence might be expected to retain its original Sr^{87} abundance.

Recent work on the Sr/Ca ratio in different faunal groups from the Permian Florena shale (Turekian, 1955) furnishes some evidence on this problem. Marked differences in the Sr/Ca ratio are reported for different faunal groups. The preservation of such great differences in the total Sr content, in a situation where opportunity for exchange is present, is not explicable if there has been extensive ionic exchange of Sr and Ca. Further evidence against extensive exchange has been cited by Patrick Hurley (Personal communication). Work in his laboratory has shown retention of most of the helium produced in ancient marble from uranium and thorium decay. This could not occur if the rock has thoroughly recrystallized with replacement of strontium in recent geological time.

Thus, although evidence against appreciable exchange of Sr may not be conclusive, the most probable conclusion is that the rate of exchange is too slow, even in terms of geologic time, to affect the Sr^{87} abundance.

In the calculation of the expected change in the Sr^{87} abundance it is assumed that, isotopically, oceanic Sr is an average of the Sr in presently exposed rocks. This appears to be an oversimplification of the actual situation. It seems likely that certain types of rocks contribute far more sedimentary material than their frequency of occurrence would indicate. Kuenen (1950, p. 392) has suggested that volcanic rocks are one of the most important sources of dissolved solids and detrital materials that enter the sea. The unusually high Ca content of sedimentary rocks derived by Poldervaart (1955) seems to confirm this suggestion. The Sr/Rb ratio in volcanic rocks (basalts and andesites) generally is much higher than the average ratio for all rocks. Thus, if calcareous materials are derived mainly from andesites and basalts, they will show a much smaller change in the Sr^{87} abundance. If about half of all sedimentary

materials are derived from basaltic rocks (the remainder coming from average surface rocks), the expected change in the Sr^{87} abundance would be roughly half of that predicted above. A further complication in the calculation results from the tacit supposition that radiogenic Sr and nonradiogenic Sr are completely mixed during weathering and transport. Actually Rb occurs almost exclusively in micas and potassium feldspars. Whereas much of the Sr will be in plagioclase and pyroxenes, the latter are much less resistant to weathering than mica and K feldspar, hence the more soluble Sr will be less radiogenic than the detrital Sr. This would tend to reduce the change of the Sr^{87} abundance with time. It seems unlikely, however, that this would account for more than a factor of two.

It is also likely that the change predicted may be too great, because of inaccuracies in the Rb and Sr content used for common igneous rocks. Recent work on the accuracy of spectrochemical techniques used in silicate analysis suggests that such analyses may be very inaccurate (Fairbairn *et al.*, 1951). In another report (Ahrens, 1954, p. 25) four different analysts report Rb and Sr contents of two standard rocks which vary by several hundred per cent. More accurate determinations of the Rb and Sr content of common rocks are needed before quantitative predictions concerning the change in Sr^{87} abundance can be made. The chemical abundances used above are also subject to errors in the estimates of the relative abundance of common rock types. The quantitative importance of such errors cannot be determined without better analytical data, but qualitatively more mafic rocks are needed in the upper crust to explain the lack of variation in the Sr^{87} abundance.

In conclusion it seems apparent that more work on the geochemistry of both Rb and Sr is necessary before the disagreement between the predicted and observed Sr^{87} abundances can be fully explained.

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